Chapter 3: Changes in Climate Extremes and their Impacts on the Natural Physical Environment

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Executive Summary

A changing climate can lead to changes in the frequency, intensity or duration of an extreme event, or result in an unprecedented, previously unobserved, extreme. As well, a weather or climate event, although not necessarily extreme in a statistical sense, still may have an extreme impact, either by crossing a critical threshold in a social, ecological or physical system, or because it occurs simultaneously with another event combined with which it leads to extreme conditions or impact. Conversely, not all extremes necessarily lead to serious impacts. Meteorological phenomena such as a tropical cyclone can have an extreme impact, depending on where and when it makes landfall, even if the specific cyclone is not extreme relative to other tropical cyclones. Changes in phenomena such as El Niño – Southern Oscillation or monsoons may affect the frequency and intensity of extremes in several regions simultaneously. [3.1]

Many weather and climate extremes are the result of natural climate variability (including phenomena such as El Niño), and natural decadal or multi-decadal variations in the climate provide the backdrop for possible anthropogenic changes. Even if there were no anthropogenic changes in climate over the next century, we can still anticipate a wide variety of natural weather and climate extremes to occur. Changes in extremes of a climate or weather variable are not always related in a simple way to changes in the mean of the same variable, and in some cases can be of opposite sign to a change in the mean of the variable (e.g., precipitation intensity may increase in some areas and seasons even if the total precipitation decreases). [3.1]

There is evidence of some changes in extremes occurring over recent decades (i.e., since 1950). It is *very likely* that there has been an overall decrease in the number of unusually cold days and nights, and an overall increase in the number of unusually warm days and nights, on the global scale, i.e., for land areas with data It is *likely* that this statement also applies at the continental scale in North America and Europe, and *very likely* that it applies in Australia. There is *medium confidence* of a warming trend in temperature extremes in Asia. There is *low confidence* in observed trends in temperature extremes in Africa and South America. It is *likely* that the number of warm spells, including heatwaves, have increased since the middle of the 20th century in many (but not all) regions. It is *likely* that there has been statistically significant increases in the number of heavy precipitation events (e.g., 95th percentile) in more regions than there has been statistically significant decreases, but there are strong regional and subregional variations in the trends. There is *low confidence* that any reported long-term increases in tropical cyclone activity are robust, after accounting for past changes in observing capabilities. There is *medium confidence* that since the 1950s some regions of the world have experienced more intense and longer droughts, in particular in southern Europe and West Africa, but also opposite trends exist e.g., in Central North America and Northwestern Australia. [3.3.1; 3.3.2; 3.4.4; 3.5.1; Table 3.2]

Our confidence in projecting changes (including the direction and magnitude in extremes) varies with the type of extreme, as well as the considered region and season, linked with the amount and quality of relevant observational data, the level of understanding of the underlying processes, and the reliability of their simulation in models. *Low confidence* in projections of a specific extreme neither implies nor excludes the possibility of changes in this extreme. [3.2.3; 3.1.5]

The following assessments of the likelihood and/or confidence of projections are generally for the end of 21st century. The reference climate period is generally 1961-1990. Projections for differing emissions scenarios generally do not strongly diverge in the coming two to three decades, but uncertainty is large over this time frame due to natural climate variability. For certain extremes (e.g., precipitation-related extremes), the overall uncertainty in projected changes by the end of the 21st century is more strongly influenced by model uncertainty than by uncertainty associated with emissions scenarios. For other extremes (in particular temperature extremes on the global scale and in most regions), the scenario uncertainty is the dominant source of uncertainty by the end of the 21st century. The provided assessments modify the uncertainty ranges from the direct evaluation of multi-model ensemble projections by taking into account the past performance of models in simulating extremes, the possibility that some important processes relevant to extremes may be missing or be poorly represented in models, and the limited number of model projections and corresponding analyses currently available of extremes. For these reasons the assessed uncertainty is generally less confident than would be the uncertainty range from the model projections alone. [3.2.3, 3.3.1, 3.3.2, Box 3.1]

Model projections of changes in temperature extremes are for substantial warming by the end of the 21st century. However, simulations of late 20th century changes in extreme temperatures suggest that although models simulate temperature extremes quite well, they may over-estimate the warming. The following assessments take this possible over-estimation into account, along with the possibility that some processes important for temperature extremes may be missing or be poorly represented in models. It is *virtually certain* that the observed increases in the warm extremes of daily temperature and decreases in their cold extremes, will continue into the future on the global scale and in most regions. It is *very likely* that the length, frequency and/or intensity of heatwaves will continue to increase over most land areas. For the SRES A2 and A1B emission scenarios, a one-in-20 year annual hottest day is *likely* to become a one-in-two year annual extreme by the end of the 21st century in most regions, except in the high latitudes of the northern hemisphere where it is *likely* to become a one-in-five year annual extreme. Moderate (cold and warm) temperature extremes on land are projected to warm faster than global annual mean temperature in many regions and seasons. A mean global warming of 2°C or 3°C is *likely* to lead to much larger increases in some temperature extremes in certain regions and seasons (with scaling factors for the SRES A2 scenario *likely* ranging between 0.5 and 2.5 for moderate seasonal extremes). Nevertheless, mean global warming does not necessarily imply warming in all regions and seasons. [3.3.1; 3.1.6; Table 3.3]

It is *likely* that the frequency of heavy precipitation (or proportion of total rainfall from heavy falls) will increase in the 21st century over many areas of the globe, in particular in the high latitudes and tropical regions, and in winter in the northern mid-latitudes. Some studies also suggest increases in heavy precipitation in some regions with projected decreases of total precipitation, such as Central Europe (*medium confidence*). For a range of emission scenarios (SRES B1/A1B/A2), a one-in-20 year annual maximum 24-hour precipitation rate is *likely* to become a one in 5- to 15-year event by the end of 21st century in many regions, and in most regions the more extreme emissions scenarios (A1B and A2) lead to a stronger projected decrease in waiting time. Nevertheless, increases or statistically non-significant changes in waiting times are projected in some regions. [3.3.2; Table 3.3]

It is *very likely* that mean sea level rise will contribute to upward trends in extreme sea levels in the future. A relatively small change in mean sea level can result in a large change in extreme sea level, in some locations. There is evidence for projected increases in wave height in some regions such as the eastern North Sea, but the small number of studies, the lack of consistency of the wind projections between models, combined with limitations in their ability to simulate extreme winds, means there is *low confidence* in these. Future negative or positive changes to significant wave height are *likely* to reflect future changes in storminess and associated patterns of wind change. There is *high confidence* that locations currently experiencing adverse impacts such as coastal erosion and inundation will continue to do so in the future due to increasing sea levels all other contributing factors being equal. [3.5.3; 3.5.4; 3.5.5]

The magnitude and even the sign of any anthropogenic influence on global patterns of floods are uncertain, and causes of regional changes in floods are complex; thus there is *low confidence* (due to *limited evidence* as well as to *low agreement of projections)* in projections of changes in flood magnitude and frequency. Nevertheless, an increase in the magnitude and/or frequency of rain-generated floods is anticipated in some catchments and regions where short-term (e.g., daily) rainfall extremes and/or long-term (e.g., monthly, wet-season total) rainfall extremes are projected to increase. Earlier spring peak flows in snowmelt and glacier-fed rivers are *very likely*. [3.5.2]

There is at most *medium confidence* in the projected changes in drought characteristics, because of inconsistent projections of the sign of changes in several regions (dependent both on model and dryness index choice). There is *medium confidence* that droughts will intensify in the 21st century in some seasons and areas (including the Mediterranean, Central Europe, Central North America, and South Africa), due to an enhanced precipitation deficit and/or evapotranspiration excess. Projections of drought intensification are stronger for the more extreme emissions scenarios (A2/A1B) than for more moderate scenarios (B1), but uncertainty between models is larger than that due to scenario choice. There is *low confidence* in projected future changes in dust activity. [3.5.1; 3.5.8; Table 3.3]

The relatively few studies of projected extreme winds, and shortcomings in the simulation of these events, mean that we have *low confidence* in projections of changes in strong winds. An exception is mean tropical cyclone maximum wind speed, which is *likely* to increase, although increases may not occur in all ocean basins. It is *likely* that tropical cyclone related rainfall rates would increase with continued warming induced by enhanced greenhouse gas concentrations, but it is *unlikely* that the global frequency of tropical cyclones would increase. A reduction in the number of mid-latitude storms averaged over each hemisphere due to future anthropogenic climate change is as *likely as not* and models show large regional changes in cyclone activity, but there is *low* confidence in the detailed geographical projections. Confidence in a projected poleward shift of mid-latitude storm tracks due to future anthropogenic forcings is *medium*. [3.3.3; 3.4.4; 3.4.5]

There is *low confidence* in projections of changes in monsoons (rainfall, circulation), even in the sign of the change, because there is little consensus in climate models regarding the sign of future change in the monsoons. Land use changes and aerosols from biomass burning appear to influence monsoons, but these effects are associated with large uncertainties. [3.4.1]

Models project a wide variety of changes in El Niño – Southern Oscillation variability and the frequency of El Niño episodes as a consequence of increased greenhouse gas concentrations, and so there is *low confidence* in projections of changes in the phenomenon. However, most models project an increase in the relative frequency of central equatorial Pacific events (which typically exhibit different patterns of climate variations than do the classical East Pacific events). [3.4.2]

There is *medium confidence* that the number of shallow landslides and debris flows from recently deglaciated terrain will increase because of higher availability of unconsolidated sediment. There is also *medium confidence* that high-mountain debris flows will begin earlier in the year because of earlier snow melt. There is *low confidence* in projected changes in the magnitude and frequency of shallow landslides in temperate and tropical regions, as these depend mainly on frequency and intensities of rainfall events and land use. It is *likely* that permafrost degradation will decrease the stability of rock slopes, though there is *low confidence* regarding future locations and times of large rock avalanches as these depend on multiple factors particularly local geological conditions. Due to *very likely* sea ice retreat, and permafrost degradation, the frequency and magnitude of the rate of Arctic coastal erosion is *likely* to increase. [3.5.6; 3.5.7]

This report identifies the most likely changes in extremes based on current knowledge. The possibility of the occurrence of low-probability high-impact scenarios associated with the crossing of poorly understood thresholds cannot be excluded given the transient and complex nature of the climate system. Non-linear feedbacks play an important role in either damping or enhancing extremes in several climate variables. [3.1.4; 3.1.7]

3.1. Weather and Climate Events Related to Disasters

3.1.1. Categories of Weather and Climate Events Discussed in this Chapter

In this chapter, we address changes in weather and climate events relevant to extreme impacts and disasters, grouped into the following categories:

- Weather and climate elements (temperature, precipitation, wind)
- Weather and climate phenomena influencing the occurrence of extreme events (monsoons, El Niño and other modes of variability, tropical and extratropical cyclones)
- Impacts on the natural physical environment (droughts, floods, extreme sea level, waves, and coastal impacts, as well as other physical impacts, including cryosphere-related impacts, landslides, and sand and dust storms)

The distinction between these three categories is somewhat arbitrary, and the categories are also related. In the case of the third category, “impacts on the natural physical environment”, a specific distinction between these events and those considered under “weather and climate elements” is that they are not induced by changes in only one of the considered weather and climate elements, but are generally the results of specific conditions in several elements, as well as of some surface properties or states. For instance, both floods and droughts are related to precipitation extremes, but are also impacted by other meteorological and surface conditions (and are thus often better viewed as compound events, see Section 3.1.3).

Another arbitrary choice made here is the separate category for phenomena that are related to weather and climate extremes, such as monsoons, El Niño, and other modes of variability. These phenomena affect the large-scale environment that, in turn, influences extremes. For instance, El Niño episodes typically see droughts in some regions with, simultaneously, heavy rains and floods occurring elsewhere. This means that all occurrences of El Niño are relevant to extremes and not only extreme El Niño episodes. A change in the frequency or nature of El Niño episodes (or in their relationships with climate in specific regions) would affect extremes in many locations simultaneously. Similarly, changes in monsoon patterns could affect several countries simultaneously. This is especially important from an international disaster perspective because coping with disasters in several regions simultaneously may be challenging (see also Section 3.1.3).

The rest of this section provides background material on the characterization and definition of extreme events, the definition and analysis of compound events, the relevance of feedbacks for extremes, the approach used for the attribution of confidence and likelihood assessments in this chapter, and the possibility of “surprises” regarding future changes in extremes. Requirements and methods for analysing changes in climate extremes are addressed in Section 3.2. Assessments regarding changes in the climate elements, phenomena and impacts considered in this chapter are provided in Sections 3.3 to 3.5. Table 3.1 provides summaries of these assessments for changes on the global scale. Tables 3.2 and 3.3 provide more regional detail on observed and projected changes in temperature and precipitation extremes. Note that impacts on ecosystems (e.g., bushfires) and human systems (e.g., urban flooding) are addressed in Chapter 4.

3.1.2. Characteristics of Weather and Climate Events Relevant to Disasters

The identification and definition of weather and climate events that are relevant from a risk management perspective is complex and depends on the stakeholders involved (Chapters 1 and 2). In this chapter, we focus on the assessment of changes in “extreme (climate or weather) events” (also referred to herein as “climate extremes”, see below), which correspond to the “hazards” discussed in Chapter 1. Hence, the present chapter does not consider the dimensions of vulnerability or exposure, which are critical in determining the human and ecosystem impacts of climate extremes (Chapters 1, 2 and 4).

The IPCC SREX defines an “extreme (climate or weather) event” as follows (SREX glossary):
“An *extreme (weather or climate) event* is generally defined as the occurrence of a value of a weather or climate variable above (or below) a threshold value near the upper (or lower) ends (“tails”) of the range of observed values of the variable. Definitions of thresholds vary, but values with less than a 5% or 1% or even lower chance of occurrence during a specified *reference* period (generally 1961-1990) are often used. Absolute thresholds (rather than these relative thresholds based on the range of observed values of a variable) can also be used to identify extreme events (e.g., specific critical temperatures for health impacts). What is called an extreme weather or climate event will vary from place to place in an absolute sense (e.g., a hot day in the tropics will be a different temperature than a hot day in mid-latitudes), and possibly in time given some adaptation from society. Extremes in some climate variables (e.g., drought or flood) may not necessarily be induced by extremes in meteorological variables (precipitation, temperature), but may be the result of an accumulation of moderate weather or climate events. Compound events (see IPCC SREX Section 3.1.3), i.e., two (even moderate) events occurring simultaneously, can lead to high impacts, even if the two single

events are not extreme per se (only their combination). Not all extreme weather and climate events necessarily have extreme impacts."

The distinction between extreme *weather* events and extreme *climate* events is not precise, but is related to their specific time scales: An extreme weather event is typically associated with changing weather patterns, i.e., within time frames of less than two weeks. An extreme climate event happens on longer time scales, i.e., from several weeks to several years or even decades. It may also be the sum of several (moderate) extreme weather events (e.g., the accumulation of above-average rainy days over a season). For simplicity, we collectively refer to both extreme weather events and extreme climate events with the term "climate extremes" in this chapter.

From the above definition, it can be seen that climate extremes can be quantitatively defined in two ways:

- Related to their probability of occurrence
- Related to a specific (possibly impact-related) threshold

The first type of definition can either be expressed with respect to given percentiles of the probability distribution functions (pdf) of the variables, or with respect to specific return frequencies (e.g., "100-year event"). Compound events can be seen as a special category of climate extremes, which result from the combination of two or more events, and which are again "extreme" either from a statistical perspective (low probability of occurrence) or associated with a specific threshold (Section 3.1.3.). These two definitions of climate extremes, probability-based or threshold-based, are not necessarily antithetic. Indeed, hazards for society and ecosystems are often extreme both from a probability and threshold perspective (e.g., a 40°C threshold for midday temperature in the mid-latitudes).

A large amount of the available scientific literature on climate extremes is based on the use of so-called "extreme indices", which can either be based on the probability of occurrence of given quantities or on threshold exceedances. Typical indices that are seen in the scientific literature include the number, percentage or fraction of cold/warm days/nights (days with maximum temperature, Tmax or minimum temperature, Tmin, below or above the 10th percentile, or the 90th percentile, generally defined with respect to the 1961-1990 reference time period). Other definitions relate to e.g., the number of days above specific absolute temperature or precipitation thresholds, or more complex definitions related to the length or persistence of climate extremes (see also below). Some advantages of using predefined extreme indices are that they allow some comparability across modelling and observational studies and across regions (although with limitations noted below). Moreover, in the case of observations, derived indices may be easier to obtain than is the case with daily temperature and precipitation data, which are not always distributed by meteorological services. Peterson and Manton (2008) discuss collaborative international efforts to monitor extremes by employing extreme indices. Typically, although not exclusively, extreme indices used in the scientific literature reflect "moderate" extremes, e.g., events occurring as often as 5% or 10% of the time. More extreme "extremes" can be better investigated using Extreme Value Theory (see below). Extreme indices are more generally defined for (daily) temperature and precipitation characteristics, and are rarely applied to other weather and climate variables, such as wind speed, humidity, or physical impacts and phenomena. Some examples are available in the literature for wind-based (Della-Marta et al., 2009) and pressure-based (Beniston, 2009a) indices, for health-relevant indices combining temperature and relative humidity characteristics (e.g., Diffenbaugh et al., 2007; Fischer and Schär, 2010), and for a range of dryness indices (see Box 3.2).

Extreme Value Theory (EVT) is another (more general) approach used for the evaluation of changes in extremes (e.g., Coles, 2001). EVT aims at deriving a probability distribution of events from the tail of a probability distribution, that is, for very low-probability (or "rare") events (typically occurring less frequently than once per year or per period of interest). Theory is used to derive a complete probability distribution for such low-probability events, which then allows also for analyzing the probability of occurrence of events which are outside of the observed data range. Two different approaches can be used to estimate the parameters for such probability distributions. In the *block maximum approach*, the probability distribution parameters are estimated for maximum values of consecutive blocks of a time series (e.g., years). In the second approach, instead of the block maxima the estimation is based on events which exceed a high threshold (*peaks over threshold* approach). Both approaches are used in climate research.

Another test for changes in extremes that has been proposed in recent years is the "iid-test" (Benestad, 2003, 2006), which investigates the occurrence probability of new record values in a time series of independent and identically distributed random variables. The iid-assumption implies stationarity of the time series. If observed records in a time series contradict the iid assumption, stationarity is not evident. For instance, Benestad (2004) reported that the monthly record temperatures at the end of the 20th century from 17 stations worldwide were higher than could be expected from stationary time series.

Beside the actual magnitude of extremes (quantified in terms of probability/return frequency or absolute threshold), other relevant aspects for the definition of climate extremes from an impact perspective include the event's duration, the spatial area affected, timing, frequency, onset date, continuity (i.e., whether there are "breaks" within a spell), and pre-conditioning (e.g., rapid transition from a slowly developing meteorological drought into an agricultural drought).

These aspects, together with seasonal variations in climate extremes, are rarely examined in climate models or observational analyses, and thus can only be partly assessed within this chapter.

To conclude, it is difficult to precisely define an extreme (e.g., Stephenson et al., 2008a) and we note limitations in the definition of both probability-based or threshold-based climate extremes and their relations to impacts, which apply independently of the chosen method of analysis:

- An event with low probability is not necessarily extreme in terms of impact
- Impact-related thresholds can vary in space and time, i.e., single absolute thresholds (e.g., a daily rainfall exceeding 25 mm or the number of frost days) will not reflect extremes in all locations and time periods (e.g., season, decade)

Orlowsky and Seneviratne (2011) illustrate these issues, by comparing projections of changes in (annual) heatwave length between 1980-2000 and 2080-2100. They compare projected changes in three different indices:

1. HWDImax or the maximum heatwave duration index (used e.g., in Frich et al., 2002; or Tebaldi et al., 2006), which is defined as the maximum length of periods of at least 5 days with a maximum air temperature (Tmax) that is 5°C higher than the climatology for that day;
2. HWDImean or the mean heatwave duration index defined as the mean length of periods of at least 5 days with Tmax that is 5°C higher than the climatology;
3. WSDI or the warm spell duration index defined as the fraction of days per year which belong to spells of at least 6 days with Tmax higher than its 90th percentile during the 1961-1990 base period (used in e.g., Alexander et al., 2006).

All three indices indicate that heatwave length is projected to increase in the models for most regions. However, the magnitude and regional patterns of the changes depend on the chosen index. The HWDImax index is not statistically robust, because it can produce many zero values for the present day in some locations, such as the tropics, where the variability of daily temperature is low (Alexander et al., 2006). To overcome this, the WSDI (percentile based threshold) has been suggested as alternative. However, this index results in very large changes in the tropics because the 90th percentile values are close to the mean value of the Tmax distributions, and thus this value is easily exceeded with the projected warming. While a typical 90th percentile value in the time frame 1961-1990 may be considered as extreme for that period, in a statistical sense, one cannot exclude the possibility that society could adapt to such temperatures, especially if it is not very far away from the mean. So the choice of using either the mean heatwave duration or the maximum heatwave duration using the same heatwave definition can lead to marked differences in projected changes. Finally, none of these indices allow us to assess changes in the intrinsic persistence of temperature extremes (i.e., the tendency for clustering of hot days, independent of changes in the number of hot days), which may be more relevant for certain impacts (e.g., Lorenz et al., 2010). Similar definition issues apply to other types of extremes, especially those characterizing dryness (see Section 3.5.1 and Box 3.2).

3.1.3. Compound (Multiple) Events

In climate science, compound events can be two or more extreme events, or combinations of extreme events with amplifying events or conditions, or combinations of events which are not themselves extremes but lead to an extreme event or impact when combined. The contributing events can be of similar (clustered multiple events) or different type(s). There are several varieties of clustered multiple events, such as tropical cyclones generated a few days apart with the same path and/or intensities, which may occur if there is a tendency for persistence in atmospheric circulation and genesis conditions. Examples of compound events resulting from events of different types are varied: for instance, high sea level coinciding with tropical cyclone landfall, or a combined risk of flooding from sea level surges and precipitation-induced high river discharge (Van den Brink et al., 2005). Compound events can even result from "contrasting extremes": e.g., the projected near-simultaneous occurrence of both droughts and heavy precipitation events in the future climate of Central Europe, (Table 3.3), or, more anecdotally, flash flooding following bushfires (due to fire-induced thunderstorms from pyrocumulus clouds (e.g., Tryhorn et al., 2008)).

Impacts on the physical environment (Section 3.5) are often the result of compound events. For instance, floods will more likely occur over saturated soils (Section 3.5.2), which means that both soil moisture status and precipitation intensity play a role. The wet soil may itself be the result of a number of above-average but not necessarily extreme precipitation events, or of enhanced snow melt associated with temperature anomalies in a given season. Similarly, droughts are both the result of pre-existing soil moisture deficits and of the accumulation of precipitation deficits and evapotranspiration excesses (Box 3.2), not all (or none) of which are necessarily extreme for a particular drought event when considered in isolation.

More generally, the following causes for a correlation between the occurrence of extremes (or their impacts) can be identified:

1. a common external forcing factor for changing the probability of the two events (e.g., regional warming, change in frequency or intensity of El Niño events)
2. mutual reinforcement of one event by the other and vice versa due to system feedbacks (Section 3.1.4)
3. conditional dependence of the occurrence or impact of one event on the occurrence of another event (e.g., extreme soil moisture levels and precipitation conditions for floods, droughts, see above)

Changes in one or more of these factors would be required, if a changing climate was to see changes in the occurrence of compound events. Unfortunately, investigation of possible changes in these factors has received little attention.

Much of the analysis of changes of extremes has, up to now, focused on individual extremes of a single variable. However, recent literature in climate research is starting to consider compound events. Compound events imply multivariate probability distributions, which can be transformed into copulas (that is, multivariate distributions, of which the marginal distributions are uniform). Schölzel and Friederichs (2008) provide an introduction to the copula approach for climate research, and an application of a Gaussian copula for multivariate hydrological extremes in France is described in Renard and Lang (2007). More general literature on multivariate extreme value theory and the statistics of multivariate extremes is provided in e.g., Coles (2001) and Beirlant et al. (2004). A more traditional approach using multivariate histograms and empirical conditional probability estimates is taken in Benestad and Haugen (2007), which analyze the relation of flood hazards with spring temperature and precipitation in Norway. In another study, Beniston (2009b) analysed changes in combined extremes (warm/dry, warm/wet, cold/dry, cold/wet) in Europe. Also, Bayesian approaches have been used to construct joint probability distributions of temperature and precipitation changes (Murphy et al., 2007; Tebaldi and Sanso, 2009).

3.1.4. Feedbacks

A special case of compound events is related to the presence of feedbacks within the climate system, i.e., mutual interaction between several climate processes, which can either lead to a damping (negative feedback) or enhancement (positive feedback) of the initial response to a given forcing. Feedbacks can play an important role in the development of extreme events, and in some cases two climate extremes can mutually strengthen one another.

One example of positive feedback between two extremes is the mutual enhancement of droughts and heatwaves in transitional regions between dry and wet climates (Seneviratne et al., 2010). This feedback has been identified as an influence on projected changes in temperature variability and heatwave occurrence in Central and Eastern Europe (Seneviratne et al., 2006a; Diffenbaugh et al., 2007), and possibly also in Britain, Eastern North America, the Amazon and East Asia (Brabson et al., 2005; Clark et al., 2006), as well as for past heatwaves and temperature extremes in Europe and the United States (Durre et al., 2000; Fischer et al., 2007a; 2007b; Hirschi et al., 2011). The two mechanisms underlying this feedback are:

1. the enhancement of soil drying with enhanced temperature (due to the higher vapour pressure deficit and its impact on evapotranspiration); and
2. the enhancement of the temperature anomalies when soil moisture deficit limits evapotranspiration and thus leads to an enhancement of sensible heat flux.

This effect is expected to play a role mostly in transitional climate regions (which are seasonally dry), since these regions are characterized by a stronger coupling of evapotranspiration with soil moisture (Koster et al., 2004b). Climate change may also lead to a shift in the location of these regions (Seneviratne et al., 2006a), and land cover may influence these feedbacks (Teuling et al., 2010). Hirschi et al. (2011) illustrate this effect for the case of Southeastern Europe, based on observational data over the time period 1961-2000. They calculate quantile regressions of the percentage of hot days (%HD) in summer in Southeastern Europe as functions of a drought index (the standardized precipitation index or SPI, see also Box 3.2) and show a pronounced widening of the %HD distribution with drier conditions. Additionally, there may also be indirect and/or non-local effects of dryness on heatwaves through e.g., changes in circulation patterns or dry air advection (e.g., Fischer et al., 2007a; Vautard et al., 2007; Haarsma et al., 2009).

Also feedbacks between trends in snow cover and changes in temperature extremes are known to be highly relevant for projections (e.g., Kharin et al., 2007; Orlowsky and Seneviratne, 2011). Both these feedbacks with soil moisture and snow affect extremes in some regions (hot extremes in transitional climate regions, and cold extremes in snow-covered regions), and thus may induce significant deviations of changes in extremes versus changes in the average climate, as also discussed in Section 3.1.6.

Other relevant feedbacks involving extreme events are those that can lead to impacts on the global climate, such as modification of land carbon uptake due to enhanced drought occurrence (e.g., Ciais et al., 2005; Friedlingstein et al., 2006; Reichstein et al., 2007), or the possible release of greenhouse gases with melting of permafrost and lakes in high-latitude regions (Davidson and Janssens, 2006; Walter et al., 2006). These aspects are not, however, considered in this chapter.

3.1.5. Confidence and Likelihood of Assessed Changes in Extremes

In this chapter, all assessments regarding past or projected changes in extremes are expressed following the new IPCC AR5 uncertainty guidance (Mastrandrea et al., 2010). The new uncertainty guidance makes a clearer distinction between confidence and likelihood but may complicate comparisons between assessments in this chapter and those in the IPCC AR4. The following procedure was followed here (see in particular Executive Summary and Tables 3.1, 3.2 and 3.3.):

- For each assessment, the confidence level for the given assessment is assessed (*low*, *medium* or *high*).
- For assessments with *high confidence*, likelihood assessments of a direction of change are also provided (*virtually certain* for 99%-100%, *very likely* for 90-100%, *likely* for 66-100%, *more likely than not* for 50-100%, *about as likely as not* for 33-66%, *unlikely* for 0-33%, *very unlikely* for 0-10%, and *extremely unlikely* for 0-1%).
- For assessments with *medium confidence*, a direction of change is provided, but without an assessment of likelihood (which we consider here to be equivalent to "more likely than not" likelihood assessments in the AR4, IPCC, 2007a).
- For assessments with *low confidence*, no direction of change is generally provided.

The confidence assessments are expert-based evaluations which consider the confidence in the tools and data basis (models, data, proxies) used to assess or project changes in a specific element, and the associated level of understanding. Examples of cases of *low confidence* for model projections are if models display poor performance in simulating the specific extreme in the present climate (see also Box 3.1), or if insufficient literature on model performance is available for the specific extreme, e.g., due to lack of observations. Similarly for observed changes, the assessment may be of *low confidence*, if the available evidence is based only on scattered data (or publications) that are insufficient to provide a robust assessment for a large region, or the observations may be of poor quality, not homogeneous, or only of an indirect nature (proxies). In cases with *low confidence* regarding past or projected changes in some extremes, we specify whether the *low confidence* is due to lack of literature, lack of evidence (data, observations), or lack of understanding. Cases of changes in extremes for which confidence in the models and data is rated as "*medium*" are those where we have some confidence in the tools and evidence available to us, but there remain substantial doubts about some aspects of the quality of these tools. **It should be noted that an assessment of *low confidence* in projections of a specific extreme neither implies nor excludes the possibility of changes in this extreme. Rather the assessment indicates low confidence in the ability to project any such changes**.

Changes (observed or projected) in some extremes are easier to assess than in others either due to the complexity of the underlying processes or to the amount of evidence available for their understanding. This results in differing levels of uncertainty in climate simulations and projections for different extremes (Box 3.1). Because of these issues, projections in some extremes are difficult or even impossible to provide, although projections in some other extremes have a high level of confidence. Overall, we can infer that **our confidence in past and future changes in extremes varies with the type of extreme, as well as the considered region and season, linked with the level of understanding and reliability of simulation of the underlying processes**. These various aspects are addressed in more detail in Box 3.1, Section 3.2 and the individual sections on specific extremes in Sections 3.3 to 3.5.

[INSERT TABLE 3.1 HERE
Table 3.1: Overview of considered extremes and summary of observed and projected changes on global scale. Regional details on observed and projected changes in temperature and precipitation extremes are provided in Tables 3.2 and 3.3.]

3.1.6. Changes in Extremes and Their Relationship to Changes in (Regional and Global) Mean Climate

Changes in extremes can be caused by changes in the mean, variance or skewness of probability distributions, or all of these. Thus a change in the frequency of occurrence of hot days (i.e., days above a certain threshold) can arise from a change in the mean daily maximum temperature, and/or from a change in the variance and/or skewness of the frequency distribution of daily maximum temperatures. If changes in the frequency of occurrence of hot days were mainly caused by changes in the mean daily maximum temperature, and changes in the shape and variability of the distribution of daily maximum temperatures were of secondary importance, then it might be reasonable to use projected changes in mean temperature to estimate how changes in extreme temperatures might change in the future. If, however, changes in the shape and variability of the frequency distribution of daily maximum temperature were important, such naive extrapolation would be less appropriate or possibly even misleading (e.g., Ballester et al., 2010).

The results of both empirical and model studies indicate that although in several situations extremes do scale closely with the mean (e.g., Griffiths et al., 2005), **there are sufficient exceptions from this that changes in the variability and shape of probability distributions of weather and climate variables need to be considered as well as changes in means, if we are to project future changes in extremes** (e.g., Caesar et al., 2006; Della-Marta et al., 2007b; Brown et al., 2008; Ballester et al., 2010; Orlowsky and Seneviratne, 2011). This appears to be especially the case for short-duration precipitation, and for temperatures at urban locations and in mid- and high-latitudes (but not all locations in these regions). In mid- and high-latitudes stronger increases (or decreases) of some extremes are generally associated with feedbacks with soil moisture or snow cover (Section 3.1.4).

An additional relevant question is the extent to which regional changes in extremes scale with changes in global mean temperature, since several studies focus on the latter (i.e., specific global mean temperature "targets", such as a 2^{o}C target, e.g., Allen et al., 2009; Meinshausen et al., 2009). As an example, Figure 3.1. displays the scaling between projected changes in the 10th and 90th percentile of Tmax on annual and seasonal (JJA, DJF) time scales with globally-averaged annual mean changes in Tmax. This scaling encompasses the effects of the scaling of regional changes with global changes, of changes in extreme quantiles with mean changes (see above), and for JJA and DJF of seasonal changes with annual mean changes (e.g., Orlowsky and Seneviratne, 2011). As seen in the figure, the scaling is rarely equal to 1, and for the 90th percentile is almost always larger than 1. This figure again highlights the particularly large projected changes in the 10th percentile Tmax in the northern high-latitude regions in winter and the 90th percentile Tmax in Southern Europe in summer (associated with snow cover and soil moisture feedbacks, respectively, Section 3.1.4), with scaling factors of up to 2.5. However, in some regions and seasons, the scaling can also be below 1 (e.g., changes in 10th percentile in JJA in the high latitudes). The lack of scaling between regional and seasonal changes in extremes and changes in means as also been highlighted in empirical studies (e.g., Caesar et al., 2006). **It should further be noted that not only do extremes not necessarily scale with mean changes, but also mean global warming does not exclude cooling in some regions and seasons**: It has for instance been recently suggested that the decrease in sea ice induced by the mean warming could induce more frequent cold winter extremes over northern continents (Petoukhov and Semenov, 2010). Also parts of Central North America and the Eastern United States present cooling trends in mean temperature and some temperature extremes in the spring to summer season (Section 3.3.1). Several mechanisms have been proposed for this cooling trend (e.g., Pan et al., 2004; Portmann et al., 2009).

[INSERT FIGURE 3.1 HERE
Figure 3.1: Scaling between globally-averaged annual mean projected change in Tmax and spatial changes in seasonal (DJF, top; JJA, bottom) changes in 10%ile (left) or 90%ile (right) of Tmax, CMIP3 projections, 2080-2100 time frame minus 1980-2000 time frame (A2 vs 20C3M). The 10%ile and 90%ile values are computed from pooling all data for the respective months in the two 20-year periods. (adapted from Orlowsky and Seneviratne, 2011)]

3.1.7. Surprises

This report focuses on the most likely changes in extremes based on current knowledge. **However, the possible future occurrence of low-probability high-impact scenarios associated with the crossing of poorly understood thresholds cannot be excluded, given the transient and complex nature of the climate system**. So, an assessment that we have *low confidence* in projections of a specific extreme, or even lack of consideration of given climate changes under the categories covered in this chapter (e.g., shutdown of meridional overturning circulation), should not be interpreted as meaning that no change is expected in this extreme or climate element (see also Section 3.1.5). Non-linear feedbacks play an important role in either damping or enhancing extremes in several climate variables (Section 3.1.4), and this can also lead to "surprises", i.e., changes in extremes greater (or less) than might be expected with a gradual warming of the climate system. Similarly, as discussed in 3.1.3, contrasting or multiple extremes can occur but our understanding of these is insufficient to provide credible comprehensive projections of risks associated with such combinations.

One aspect that we do not address in this chapter is the existence of possible tipping points in the climate system (e.g., Lenton et al., 2008; Scheffer et al., 2009), that is, the risks of abrupt, possibly irreversible changes in the climate system. Scheffer et al. (2009) illustrate the possible equilibrium responses of a system to forcing. In the case of a linear response only a large forcing can lead to a major state change in the system, and this change is potentially reversible. However, in the presence of a critical threshold even a small change in forcing can lead to a major (reversible) change in the system. For systems with critical bifurcations in the equilibrium state function two alternative stable conditions may exist, whereby an induced change may be (temporarily or permanently) irreversible. Such critical transitions within the climate system represent typical low-probability high-risks scenarios, which cannot be excluded at present. Lenton et al. (2008) provided a recent review on tipping elements within the climate system, i.e., sub-systems of the Earth system that are at least sub-continental in scale and which entail a tipping point. They identified the following tipping elements for the Earth's climate system: the Greenland ice sheet, the Arctic summer sea-ice, the West Antarctic ice sheet, the Atlantic thermohaline circulation, El Niño – Southern Oscillation, the Indian summer monsoon, the Sahara/Sahel and West African monsoon, the Amazon rainforest, and the boreal forest. Some of these would be especially relevant to certain extremes (e.g., El Niño – Southern Oscillation for drought, and the ice sheets for sea level extremes), or are induced by changes in extremes (e.g., Amazon rainforest die-back induced by drought). For some of the tipping elements, the existence of bistability can be inferred from paleo records. There is often a lack of agreement between models regarding these high-risk low-probability scenarios, for instance regarding a possible die-back of the Amazon rainforest (Friedlingstein et al., 2006) or the risk of an actual shutdown of the Atlantic thermohaline circulation (e.g., Lenton et al., 2008). Scheffer et al. (2009) have recently suggested the existence of early signs for approaching critical transitions.

3.2. Requirements and Methods for Analysing Changes in Extremes

3.2.1. Observed Changes

Sections 3.3 to 3.5 of this Chapter provide assessments of the literature regarding changes in extremes in the observed record published mainly since the AR4. Summaries of these assessments are provided in Table 3.1. Overviews of observed regional changes in temperature and precipitation extremes are provided in Table 3.2. In this sub-section issues are discussed related to the data and observations used to examine observed changes in extremes.

Issues with data availability are especially critical when examining changes in extremes of given climate variables (Nicholls, 1995). Indeed, the more rare the event, the more difficult it is to identify long-term changes, simply because there are fewer cases to evaluate (Frei and Schär, 2001; Klein Tank and Können, 2003). Identification of changes in extremes is also dependent on the analysis technique employed (Zhang et al., 2004b; Trömel and Schönwiese, 2005). Another important criterion constraining data availability for the analysis of extremes is the respective time scale on which they occur (Sections 3.1.2), since this determines the required temporal resolution for their assessment (e.g., heavy hourly or daily precipitation versus multi-year drought). Longer time resolution data (e.g., monthly, seasonal, and annual values) for temperature and precipitation are available for most parts of the world starting late in the 19th to early 20th century, and allow analysis of (meteorological) drought (see Box 3.2) and unusually wet periods on the order of a month or longer. To examine changes in extremes occurring on short time scales, particularly of climate elements such as temperature and precipitation (or wind), normally requires the use of high-temporal resolution data, such as daily or sub-daily observations, which are generally either not available, or available only since the middle of the 20th century and in many regions only from as recently as 1970. Even where sufficient data are available, several problems can still limit their analysis. First, although the situation is changing (especially for the situation with respect to “extreme indices”, Section 3.1.2), many countries still do not freely distribute their higher temporal resolution data. Second, there can be issues with the quality of measurements. A third important issue is climate data homogeneity (see below). These and other issues are discussed in detail in the AR4 (Trenberth et al., 2007). For instance, the temperature and precipitation stations considered in the daily dataset used in Alexander et al. (2006) are not globally uniform, and measurements are in particular found to be lacking in Northern South America, Africa, and part of Australia. The other commonly used dataset from Caesar et al. (2006; used e.g., in Brown et al., 2008) has additional data gaps in most of South America, Africa, Eastern Europe, Mexico, the Middle East, India, and Southeast Asia. Also the study by Vose et al. (2005) has data gaps in South America, Africa and India. It should be further noted that the regions with data coverage do not all have the same density of stations (Alexander et al., 2006; Caesar et al., 2006). While some studies are available on a country- or regional basis for areas not covered in global studies, nevertheless lack of data leads to limitations in our ability to assess observed changes in climate extremes for many regions

Whether or not climate data are homogeneous is of strong relevance for the results of an analysis of extremes. Data are defined as homogeneous when the variations and trends in a climate time series are due solely to variability and changes in the climate system. Some meteorological elements are especially vulnerable to uncertainties caused by even small changes in the exposure of the measuring equipment. For instance, erection of a small building or changes in vegetative cover near the measuring equipment can produce a bias in wind measurements (Wan et al., 2010). When a change occurs it can result in either a discontinuity in the time series (slight jump) or a more gradual change that can manifest itself as a false trend (Menne and Williams Jr., 2009), both of which can impact on whether a particular observation exceeds a threshold. Homogeneity detection and data adjustments have been implemented for longer averaging periods (e.g., monthly, seasonal, annual); however techniques applicable to daily and sub-daily data are only now being developed (e.g., Vincent et al., 2002; Della-Marta and Wanner, 2006), and have not been widely implemented. Homogeneity issues also affect the monitoring of other meteorological and climate variables, for which further and more severe limitations also can exist. This is in particular the case regarding measurements of wind and relative humidity, and data required for the analysis of weather and climate phenomena (tornadoes, extra-tropical and tropical cyclones, Section 3.4), as well as impacts on the physical environment (e.g., droughts, floods, cryosphere impacts, Section 3.5).

Thunderstorms, tornadoes and related phenomena are not well observed in many parts of the world. Tornado occurrence since 1950 in the USA., for instance, displays an increasing trend that mainly reflects increased population density and increased numbers of people in remote areas (Trenberth et al., 2007; Kunkel et al., 2008). Such trends increase the likelihood that a tornado would be observed. A similar problem occurs with thunderstorms. Changes in reporting practices, increased population density and even changes in the ambient noise level at an observing station all have led to inconsistencies in the observed record of thunderstorms.

Studies examining changes in extratropical cyclones, which focus on changes in storm track location, intensities and frequency, are limited in time due to a lack of suitable data prior to about 1950. Most of these studies have relied on model-based reanalyses that also incorporate observations into a hybrid model-observational data set. However, reanalyses can have homogeneity problems due to changes in the amount and type of data being assimilated, such as the introduction of satellite data in the late 1970s and other observing system changes (Trenberth et al., 2001; Bengtsson et al., 2004). Recent efforts in reanalysis have attempted to produce more homogeneous reanalyses that show promise for examining changes in extratropical cyclones and other climate features (Compo et al., 2006). Results, however, are strongly dependent on the reanalysis and cyclone tracking techniques used (Ulbrich et al., 2009).

The robustness of analyses of observed changes in tropical cyclones has been hampered by a number of issues with the historical record. One of the major issues is the heterogeneity introduced by changing technology and reporting protocols within the responsible agencies (e.g., Landsea et al., 2004). Further heterogeneity is introduced when records from multiple ocean basins are combined to explore global trends, because data quality and reporting protocols vary substantially between agencies (Knapp and Kruk, 2010). Much like other weather and climate observations, tropical cyclone observations are taken to support short-term forecasting needs. Improvements in observing techniques are often implemented without any overlap or calibration against existing methods to document the impact of the changes on the climate record. Additionally, advances in technology have enabled better and more complete observations. For example, the introduction of aircraft reconnaissance in some basins in the 1940s and satellite data in the 1960s had a profound effect on our ability to accurately identify and measure tropical cyclones, particularly those that never encountered land or a ship. While aircraft reconnaissance programs have continued in the North Atlantic, they were terminated in the Western Pacific in 1987. The introduction of geostationary satellite imagery in the 1970s, and the introduction (and subsequent improvement) of new tropical cyclone analysis methods (such as the Dvorak technique for estimating storm intensity), further compromises the homogeneity of historical records of tropical cyclone activity.

Regarding impacts to the physical environment, soil moisture is a key variable for which data sets are extremely scarce (e.g., Robock et al., 2000; Seneviratne et al., 2010). This represents a critical issue for the validation and correct representation of (agricultural as well as hydrological) drought mechanisms in climate, land surface and hydrological models, and the monitoring of on-going changes in regional terrestrial water storage. As a consequence, these need to be inferred from simple climate indices or model-based approaches (Box 3.2). Such estimates rely in large part on precipitation observations, which have, however, inadequate spatial coverage for these applications in many regions of the world (e.g., Oki et al., 1999; Fekete et al., 2004; Koster et al., 2004a). Similarly, runoff observations are not globally available, which results in significant uncertainties in the closing of the global and some regional water budgets (Legates et al., 2005; Peel and McMahon, 2006; Dai et al., 2009; Teuling et al., 2009), as well as for the global analysis of changes in the occurrence of floods. Additionally, ground observations of snow, which are lacking in several regions, are important for the investigation of several physical impacts, in particular those related to the cryosphere and runoff generation (e.g., Essery et al., 2009; Rott et al., 2010).

All of the above-mentioned issues lead to uncertainties in observed trends in extremes. In many instances, great care has been taken to develop procedures to improve the data which in turn helps to reduce uncertainty and progress has been made in the last 15 years (e.g., Caesar et al., 2006; Brown et al., 2008). As a consequence, more complete and homogenous information about changes is now available for at least some variables and regions (Nicholls and Alexander, 2007; Peterson and Manton, 2008). For instance, the development of global data bases of daily temperature and precipitation covering up to 70% of the global land area, has allowed robust analyses of extremes (c.f., Alexander et al., 2006). In addition, analyses of temperature and precipitation extremes using higher temporal resolution data, such as that available in the Global Historical Climatology Network-Daily data set (Durre et al., 2008) have also proven robust on both a global (Alexander et al., 2006) and regional basis (Sections 3.3.1 and 3.3.2). Nonetheless, as highlighted above, for many extremes, data remain sparse and problematic resulting in less ability to establish changes particularly on a global basis.

[INSERT FIGURE 3.2 HERE
Figure 3.2: Definitions of regions used in Tables 3.2 and 3.3.]

[INSERT TABLE 3.2 HERE
Table 3.2: Regional observed changes in temperature and precipitation extremes, including dryness. See Figure 3.2 for definitions of regions.]

3.2.2. The Causes Behind the Changes

This section discusses the main requirements, approaches, and considerations for the attribution of causes for observed changes in extremes. In Sections 3.3. to 3.5, the causes for observed changes in specific extremes are assessed. A global summary of these assessments is provided in Table 3.1

3.2.2.1. Why Extremes Change and What are the Possible Causes

Climate variations and change are induced by variability internal to the climate system, and changes in external forcings, which include natural external forcings such as changes in solar irradiance and volcanism, and anthropogenic forcings such as increased greenhouse gas emissions principally due to the burning of fossil fuels, and land use and land cover changes. The mean state, extremes, and variability are all related aspects of the climate, so changes that affect the mean climate would in general result in changes in extremes. For this reason, we provide in section 3.2.2.2 a brief overview of human-induced changes in the mean climate to aid the understanding of changes in extremes as the literature directly addressing the causes of changes in extremes is quite limited.

3.2.2.2. Human-Induced Changes in the Mean Climate that Affect Extremes

The occurrence of extremes is usually the result of multiple factors, which can act either on the large scale or on the regional (and local) scale. Some relevant large-scale impacts of external forcings affecting extremes include the overall increases in temperature induced by changes in radiation, the enhanced humidity content of the atmosphere, the increased land-sea contrast in temperatures, which can, e.g., affect circulation patterns and in particular monsoons. On the regional and local scales, additional processes can modulate the overall changes in extremes, including regional feedbacks, in particular linked to land-atmosphere interactions with e.g., soil moisture or snow (e.g., Section 3.1.4). This section briefly reviews the current understanding of the causes (i.e., in the sense of attribution to either external forcing or internal climate variability) of large-scale (and some regional) changes in the mean climate that are of relevance to extreme events, to the extent that they have been considered in detection and attribution studies.

Regarding observed increases in global average annual mean surface temperatures in the second half of the 20th century, we base our analysis on the following AR4 assessment (IPCC, 2007a): Most of the observed increase in global average temperatures is *very likely* due to the observed increase in anthropogenic greenhouse gas concentrations. Greenhouse gas forcing alone would *likely* have resulted in a greater warming than observed if there had not been an offsetting cooling effect from aerosol and other forcings. It is *extremely unlikely* (<5%) that the global pattern of warming can be explained without external forcing, and *very unlikely* that it is due to known natural external causes alone. Anthropogenically-forced warming over the second half of the 20th century has also been detected in all continents, in addition to the global-scale attribution (Hegerl et al., 2007; Gillett et al., 2008b).

Overall, attribution at scales smaller than continental, with limited exceptions (below), has still not yet been established primarily due to the low signal-to-noise ratio and the difficulties of separately attributing effects of the wider range of possible driving processes (either attributable to external forcing or internal climate variability) at these scales. Moreover, averaging over smaller regions reduces the internal variability less than does averaging over large regions. In addition, the small-scale details of external forcing, and the responses simulated by models are less credible than large-scale features: For instance, temperature changes are poorly simulated by models in some regions and seasons (Dean and Stott, 2009; van Oldenborgh et al., 2009). Also the inclusion of additional forcing factors, such as land-use change and aerosols that are likely more important at regional scales, remains a challenge (Lohmann and Feichter, 2007; Pitman et al., 2009; Rotstayn et al., 2009). Nonetheless, recent work has expanded the literature and showed more evidence of detection of an anthropogenic influence at increasingly smaller spatial scales and for seasonal averages (Stott et al., 2010). For instance, Min and Hense (2007) found that anthropogenic forcing as opposed to alternative explanation such as natural external forcing or internal variability was required for most continent-season temperature changes in multi-model estimates from the CMIP-3 ensemble to best match the observed changes. An anthropogenic signal was detected in 20th century summer temperatures in each of 14 Northern Hemispheric sub-continental regions except central North America, although the results were more uncertain when anthropogenic and natural signals were considered together (Jones et al., 2008). An anthropogenic signal has also been detected in multi-decadal trends of a U.S. climate extreme index (Burkholder and Karoly, 2007), in the hydrological cycle of the western United States (Barnett et al., 2008), in New Zealand temperatures (Dean and Stott, 2009), and in Europe (Christidis et al., 2011).

One of the significant advances since AR4 is the emerging evidence of human influence on global atmospheric moisture content and precipitation. According to the Clausius-Clapeyron relationship, the saturation vapor pressure increases exponentially with temperature. Since moisture condenses out of supersaturated air, it is physically plausible that the distribution of relative humidity would remain roughly constant under climate change. This means that specific humidity increases about 7% for a one degree increase in temperature. Indeed, observations indicate significant increases between 1973 and 2003 in global surface specific humidity but not in relative humidity (Willett et al., 2008), and at the largest spatial-temporal scales moistening is close to the Clausius-Clapeyron scaling of the saturated specific humidity (~7%/K, Willett et al., 2010), though relative humidity over low- and mid-latitude land areas decreased over a 10-year period prior to 2008 possibly due to slower temperature increase in the oceans than over the land (Simmons et al., 2010). Anthropogenic influence has been detected in the global surface specific humidity for 1973–2003 (Willett et al., 2007), and in lower tropospheric moisture content over the 1988–2006 period (Santer et al., 2007).

The increase in the atmospheric moisture content would be expected to lead to an increase in extreme precipitation. Measurements in the Netherlands suggest that hourly precipitation extremes may in some cases increase more strongly with temperature (twice as fast) than would be assumed from the Clausius-Clapeyron relationship alone (Lenderink and Van Meijgaard, 2008; Haerter and Berg, 2009; Lenderink and van Meijgaard, 2009). The influence of anthropogenic forcing has been detected in the pattern of land precipitation trends though the model-simulated magnitude of changes is smaller than that observed (Zhang et al., 2007a). Because models do not simulate exactly the same spatial pattern of precipitation trends, the simple averaging of those patterns from model simulation in Zhang et al. (2007a) would tend to reduce the model signal. The influence of anthropogenic greenhouse gases and sulphate aerosols on changes in precipitation over high-latitude land areas north of 55°N has also been detected (Min et al., 2008). Detection is possible here, despite limited data coverage, in part because the response to forcing is relatively strong in the region, and because internal variability is low in this region.

3.2.2.3. How to Attribute a Change in Extremes to Causes

The guidance paper on detection and attribution (Hegerl et al., 2010) from the joint IPCC WGI/WGII expert meeting on detection and attribution (Sept. 14-16, 2009) provides detailed guidance on the procedures that include two main approaches to attribute a change in climate to causes. One is single-step attribution that involves assessments that attribute an observed change within a system to an external forcing based on explicitly modelling the response of the variable to the external forcings. The alternate procedure is multi-step attribution that combines an assessment that attributes an observed change in a variable of interest to a change in climate, with a separate assessment that attributes the change in climate to external forcings. Attribution of changes in climate extremes has some unique issues. Observed data are limited in both quantity and quality (Section 3.2.1), resulting in uncertainty in the estimation of past changes; the signal-to-noise ratio may be low for many variables and insufficient data may be available to detect such weak signals. On the other hand, GCMs have several issues simulating extremes (Section 3.2.3).

Single-step attribution based on optimal detection and attribution (e.g., Hegerl et al., 2007) can in principle be applied to climate extremes. However, the difference in statistical properties between mean values and extremes needs to be carefully addressed (e.g., Zwiers et al., 2011; see also Section 3.1.6). Post-processing of climate model simulations to derive a quantity of interest that is not explicitly simulated by the models, by applying empirical methods or physically-based models to the outputs from the climate models, may make it possible to directly compare observed extremes with climate model results. For example, sea level pressure simulated by multiple GCMs has been used to derive geostrophic wind to represent atmospheric storminess and to derive significant wave height on the oceans for the detection of external influence on trends in atmospheric storminess and northern oceans wave heights (Wang et al., 2009d). GCM-simulated precipitation and temperature have been downscaled as input to hydrological and snow depth models to infer past and future changes in temperature, timing of the peak flow, and snow water equivalent for the western U.S., and this enabled a detection and attribution analysis on human-induced changes in these variables (Barnett et al., 2008)

If a single-step attribution of causes to effects on extremes or physical impacts of extremes is not feasible, it might be feasible to conduct a multiple-step attribution. The assessment would then need to be based on evidence not directly derived from model simulations, physical understanding and expert judgement, or their combination. For instance, in the northern high latitude regions, spring temperature has increased, and the timing of spring peak floods of snowmelt rivers has shifted towards earlier dates (Zhang et al., 2001; Regonda et al., 2005). A change in streamflow may be attributable to external influence if streamflow regime change can be attributed to a spring temperature increase and if the spring temperature increase can be attributed to external forcings. In such a case, it may not be possible to quantify the magnitude of the effect of external forcing on flow regime change because a direct link between the two has not been established, so the confidence in the overall assessment would be similar to, or weaker than, the lower confidence in the two steps in the assessment. The physical understanding that snow melts earlier as spring temperature increases enhances our confidence in the assessments. In cases where the underlying physical mechanisms are less certain, such as those linking tropical cyclones and sea surface temperature (see section 3.4.4), the confidence in multi-step attribution can be severely undermined. A necessary condition for multi-step attribution is to establish the chain of mechanisms responsible for the specific extremes being considered. Physically-based process studies and sensitivity experiments that help the physical understanding can play an important role in such cases (e.g., Findell and Delworth, 2005; Seneviratne et al., 2006a; Haarsma et al., 2009).

Extreme events are by definition rare, which means that there are also few data available to make an assessment (Section 3.2.1). When a rare and catastrophic meteorological extreme event occurs, a question that is often posed is whether such an event is due to anthropogenic influence. Because it is very difficult to rule out the occurrence of low probability events in an unchanged climate and the occurrence of such events usually involves multiple factors, it is very difficult to attribute an individual event to external forcing (Allen, 2003; Hegerl et al., 2007, see also FAQ 3.2). However, in this case, it may be possible to estimate the influence of external forcing on the likelihood of such an event occurring (e.g., Stott et al., 2004; Pall et al., 2011).

3.2.3. *Projected Long-Term Changes and Uncertainties*

In this sub-section we discuss the requirements and methods used for preparing climate change projections, with a clear focus on projections of extremes and the associated uncertainties. Much of the discussion is based closely on AR4 (Christensen et al., 2007) with consideration of some additional issues relevant to projections of extremes in the context of risk and disaster management. More detailed assessment of projections for specific extremes is provided in Sections 3.3 to 3.5. Summaries of these assessments are provided in Table 3.1. Overviews of projected regional changes in temperature and precipitation extremes are provided in Table 3.3.

3.2.3.1. Information Sources for Climate Change Projections

Work on the construction, assessment and communication of climate change projections, including regional projections and of extremes, typically draws on information from four sources: Atmosphere-Ocean General Circulation Model (AOGCM) simulations, also referred to as General Circulation Models (GCMs); downscaling of GCM-simulated data using techniques to enhance regional detail; physical understanding of the processes governing regional responses; and recent historical climate change. At the time of the AR4, GCMs were the main source of globally-available regional information on the range of possible future climates including extremes (Christensen et al., 2007). This is still the case for many regions, as can be seen in Table 3.3.

State-of-the-art GCMs show significant and improving skill in representing many important average climate features, and even essential aspects of many of the patterns of climate variability observed across a range of time scales. In particular, the AR4 demonstrated that global statistics of extreme events for present day climate are surprisingly well simulated by current GCMs considering their resolution and large-scale systematic errors (Randall et al., 2007). This makes them ‘fit for purpose’ for many applications. However, when we wish to project climate and weather extremes, not all atmospheric phenomena potentially of relevance can be realistically simulated using these global models. Much of the signal from climate change simulations is sensitive to model parameterization schemes (e.g., radiation, land-surface, and cloud schemes). Furthermore, the assessment of climate model performance with respect to extremes (summarised in Sections 3.3 to 3.5 for specific extremes), particularly at the regional or local scale, is still limited by the fact that the very rarity of extreme events makes statistical evaluation of model performance less robust than is the case for average climate. Also, evaluation is still hampered by incomplete data on the historical frequency and severity of extremes, particularly for variables other than temperature and precipitation (Section 3.2.1).

The development of projections of extreme events has provided one of the motivations for the development of regionalisation or downscaling techniques (Carter et al., 2007). Downscaling techniques have been specifically developed for the study of regional- and local-scale climate change, to simulate weather and climate at finer spatial resolutions than is possible with GCMs – a step which is particularly relevant for many extremes given their spatial scale. They are, nonetheless, constrained by the reliability of large-scale information coming from the GCMs. Recent advances in downscaling for extremes are discussed below. However, as global models continue to develop, and their spatial resolution as well as their complexity continues to improve, they will become increasingly useful for investigating important smaller-scale features, including changes in extreme weather events, and further improvements in regional-scale representation are expected with increased computing power.

There are two main downscaling approaches, dynamical and statistical (Christensen et al., 2007). The most common approach to dynamical downscaling uses high-resolution regional climate models (RCMs), currently at scales of 20km-50 km, but in some cases down to 10-15 km (e.g., Dankers et al., 2007), to represent regional sub-domains, using either observed (reanalysis) or lower-resolution GCM data to provide their boundary conditions. Using non-hydrostatic mesoscale models, applications at 1-5 km resolution are also possible for shorter periods (typically a few months, a few full years at most) – a scale at which clouds and convection can be resolved and the diurnal cycle tends to be better resolved (e.g., Grell et al., 2000; Hay et al., 2006; Hohenegger et al., 2008). Less-commonly used approaches to dynamical downscaling involve the use of stretched-grid (variable resolution) models and high-resolution ‘time-slice’ models (e.g., Cubasch et al., 1995; Gibelin and Deque, 2003; Coppola and Giorgi, 2005; CCSP, 2008). The latter have been run at 20 km globally in the case of the ‘super-high’ resolution simulations (Kamiguchi et al., 2006; Kitoh et al., 2009; Kim et al., 2010).

The main advantage of dynamical downscaling is its potential for capturing mesoscale nonlinear effects and providing information for many climate variables at a relatively high spatial resolution, while ensuring that such information is internally consistent within the physical constraints of the model. For many users, the main drawbacks of dynamical models for downscaling are their computational cost and that they do not provide information at the point (i.e., weather station) scale (a scale at which the RCM parameterizations would not work). RCMs provide area-averaged precipitation which means a tendency to more days of light precipitation (Frei et al., 2003; Barring et al., 2006) and reduced magnitude of extremes (Chen and Knutson, 2008; Haylock et al., 2008) compared with point values. Other concerns with RCMs are that they may involve different parameterization schemes to the driving models and most currently do not include coupling between ocean and atmosphere (Wang et al., 2004). Moreover, questions remain about RCM domain size and the extent to which RCM features, like biases, are inherited from the driving model, and the implications of these issues for the credibility of the RCM projections (Wang et al., 2004; Laprise et al., 2008).

Statistical downscaling methods use relationships between the large-scale circulation (predictands) and local-scale surface variables (predictors) that have been derived from observed data, and apply these to climate model data (Christensen et al., 2007). They may also include weather generators which provide the basis for a number of recently-developed user tools that can be used to assess changes in extreme events (Kilsby et al., 2007; Burton et al., 2008; Qian et al., 2008; Semenov, 2008). Statistical downscaling has been demonstrated to have potential in a number of different regions including Europe (e.g., Schmidli et al., 2007), Africa (e.g., Hewitson and Crane, 2006), Australia (e.g., Timbal et al., 2008; Timbal et al., 2009), South America (e.g., D'Onofrio et al., 2010) and North America (e.g., Vrac et al., 2007; Dibike et al., 2008). Statistical downscaling methods have the advantage to users of being computationally inexpensive, able to access finer spatial scales than dynamical methods and applicable to parameters that cannot be directly obtained from the RCM outputs. Seasonal indices of extremes can, for example, be simulated directly without having to first produce daily time series (Haylock et al., 2006a) or distribution functions of extremes can be simulated (Benestad, 2007). However, they require observational data at the desired scale (e.g., the point or station scale) for a long enough period to allow the model to be well trained and validated, and in some methods, can lack coherency among multiple climate variables and/or multiple sites. In the case of downscaling extremes, one specific disadvantage of some analog statistical methods is that they cannot produce events greater in magnitude than have been observed before (Timbal et al., 2009). In addition, both present-day performance and the projected climate change can be very sensitive to the choice of predictors (Charles et al., 1999; Hewitson and Crane, 2006). Finally, a potential limitation of statistical downscaling methods is that their calibration is necessarily based on present (and past) climate, and they may thus not be able to capture changes in extremes that are induced by mechanistic changes in regional (or global) climate. There have been few systematic inter-comparisons of dynamical and statistical downscaling approaches focusing on extremes (Fowler et al., 2007a). Two examples focus on extreme precipitation for the UK (Haylock et al., 2006a) and the Alps (Schmidli et al., 2007), respectively.

In terms of temporal resolution, while GCMs and RCMs operate at sub-daily timesteps, output is rarely archived at six-hourly or shorter temporal resolutions. Where limited studies have been undertaken of RCMs, there is evidence that at the typically used spatial resolutions (i.e., non-cloud/convection resolving scales) they do not adequately represent sub-daily precipitation and the diurnal cycle of convection (Gutowski et al., 2003; Brockhaus et al., 2008; Lenderink and Van Meijgaard, 2008). Development of sub-daily statistical downscaling methods is constrained by the availability of long observed time series for calibration and validation and this approach is not currently widely used for climate change applications, although some weather generators, for example, do provide hourly information (Maraun et al., 2010).

For reasons of space, it is not possible in this chapter to provide assessments of projected changes in extremes at scale finer than regional (Table 3.3.). Several countries, in particular in Europe, have, however, developed their own national projections, including information about extremes, and a range of other high-resolution information and tools are available from national weather and hydrological services and academic institutions to assist users and decision makers.

3.2.3.2. Uncertainty Sources in Climate Change Projections

Uncertainty in climate change projections arises at each of the steps involved in their preparation: determination of greenhouse gas and aerosol emissions, concentrations of radiatively active species, radiative forcing, and climate response including downscaling. At each step, uncertainty in the estimation of the true “signal” of climate change is introduced by both errors in the model representation of Earth system processes and by internal climate variability.

As was noted in Section 3.2.3.1, most shortcomings in GCMs and in RCMs result from the fact that many important small-scale processes (e.g., representations of clouds, convection, land-surface processes) are not represented explicitly (Randall et al., 2007). Some processes – particularly those involving feedbacks (Section 3.1.4), and this is especially the case for climate extremes and associated impacts - are still poorly represented and/or understood (e.g., land-atmosphere interactions, stratospheric processes, blocking dynamics) despite some improvements in the simulations of others (see Box 3.1 and below). Therefore, limitations in computing power and in the scientific understanding of some physical processes, currently restrict further global and regional model improvements. In addition, uncertainty due to structural or parameter errors in GCMs propagates directly from global model simulations as input to downscaling models and thus to downscaled information.

These problems limit quantitative assessments of the magnitude and timing, as well as regional details, of some aspects of projected climate change. For instance, even atmospheric models with approximately 20 km horizontal resolution still do not resolve the atmospheric processes sufficiently finely to simulate the high wind speeds and low pressure centres of the most intense hurricanes (Gutowski et al., 2008a). Realistically capturing details of such intense hurricanes, such as the inner eyewall structure, would require models with 1 km horizontal resolution, far beyond the capabilities of current GCMs and of most current RCMs (and even numerical weather prediction models). Extremes may also be impacted by mesoscale circulations that GCMs and even current RCMs cannot resolve, such as low-level jets and their coupling with intense precipitation (Anderson et al., 2003; Menendez et al., 2010). Another issue with small-scale processes is the lack of relevant observations, such as is the case e.g., with soil moisture and vegetation processes (Section 3.2.1.) and relevant parameters (e.g., maps of soil types, c.f. Seneviratne et al., 2006b; Anders and Rockel, 2009).

Since many extreme events occur at rather small temporal and spatial scales, where climate simulation skill is currently limited and local conditions are highly variable, projections of future changes cannot always be made with a high level of confidence (Easterling et al., 2008). The credibility in projections of changes in extremes varies with extreme type, season, and geographical region (Box 3.1). Confidence and credibility in projected changes in extremes increase when the physical mechanisms producing extremes in models are considered reliable, such as increases in specific humidity in the case of the projected increase in the proportion of summer precipitation falling as intense events in Central Europe (Kendon et al., 2010). The ability of a model to capture the full distribution of variables – not just the mean – together with long-term trends in extremes, implies that some of the processes relevant to a future warming world may be captured (van Oldenborgh et al., 2005; Alexander and Arblaster, 2009). It should nonetheless be stressed that physical consistency of simulations with observed behaviour provides only necessary and not sufficient evidence for credible projections (Gutowski et al., 2008a).

While downscaling provides more spatial detail, the added value of this step needs to be assessed (Benestad et al., 2007; Laprise et al., 2008), keeping in mind that an overfitted model may perform well for present climate but will not be credible for future projections. Spatial inhomogeneity of both land-use/land-cover and aerosol forcing, adds to regional uncertainty. This means that the factors inducing uncertainty in the projections of extremes in different regions may differ considerably. Specific issues inducing uncertainties in RCM projections are the interactions with the driving GCM, especially in terms of biases and climate-change signal (e.g., Déqué et al., 2007; de Elía et al., 2008; Laprise et al., 2008; Kjellstrom and Lind, 2009) and the choice of regional domain (Wang et al., 2004; Laprise et al., 2008). In the case of statistical downscaling, uncertainties are induced by the choice of domain size (Benestad, 2001) together with the choice of predictors themselves (Charles et al., 1999; Hewitson and Crane, 2006) and the underlying assumption of stationarity (Raje and Mujumdar, 2010). For both dynamical and statistical downscaling, uncertainties are also inherited from the GCMs that provide the large-scale changes driving the downscaling models.

For many user-driven applications, impact models need to be included as an additional step for projections (e.g., hydrological or ecosystem models). Because of the mentioned issues of scale discrepancies and overall biases, it is necessary to bias correct RCM data before input to some impacts models (i.e., to bring the statistical properties of present-day simulations in line with observations and to use this information to correct projections). A number of bias correction methods, including quantile mapping and gamma transform, have recently been developed and indicate promising skill for extremes of daily precipitation (Piani et al., 2010; Themeßl et al., 2011).

In conclusion, it has been recommended (Knutti et al., 2010b) that the following four factors should be considered in assessing the likely future climate change in a region: historical change, process change, global climate change projected by GCMs, and downscaled projected change. Consistency and comprehensiveness of the physical and dynamical basis of the projected climate response across models and methods should be evaluated. Moreover, model evaluation, detection and attribution, observations and projections are intimately linked, and assessments of climate projections would benefit from a tighter integration of these topics (Knutti et al., 2010a). How to address this issue in the context of extremes is discussed further in the next section.

3.2.3.3. Ways of Exploring and Quantifying Uncertainties

Uncertainties can be explored, and quantified to some extent, through the combined use of observations, process understanding, a hierarchy of climate models, and ensemble simulations. Ensembles of model simulations represent a fundamental resource for studying the possible range of plausible climate responses to a given forcing (Meehl et al., 2007b; Randall et al., 2007). Such ensembles can be generated either by (i) collecting results from a range of models from different modelling centres (multi-model ensembles), to include the impact of structural model differences, (ii) by generating simulations with different initial conditions (intra-model ensembles) to characterize the uncertainties due to internal climate variability, or (iii) varying multiple internal model parameters within plausible ranges (perturbed and stochastic physics ensembles), with both (ii) and (iii) aiming to produce a more systematic estimate of single model uncertainty (Knutti et al., 2010b).

Many of the global models utilized for the AR4 were integrated as ensembles, permitting more robust statistical analysis than is possible if a model is only integrated to produce a single projection. Thus the GCM simulations reflect both inter- and intra-model variability. In advance of AR4, coordinated climate change experiments were undertaken which provided information from 23 models from around the world (Meehl et al., 2007a). The simulations (referred to henceforth as the CMIP3 MME – Coupled Model Intercomparison Project 3 multi-model ensemble) were made available at the Program for Climate Model Diagnosis and Intercomparison (PCMDI, http://www-pcmdi.llnl.gov/ipcc/about_ipcc.php). However, the higher temporal resolution (i.e., daily) data necessary to analyze most extreme events were quite incomplete in the archive, with only four models providing daily averaged output with ensemble sizes greater than three realizations and many models not included at all. GCMs are expensive to run, and thus a compromise is needed between the number of models, number of simulations and the complexity of the models (Knutti, 2010). The Coupled Model Intercomparison Project Phase 5 multi-model ensemble (CMIP5 MME) is currently being implemented and will provide a new framework for coordinated climate change experiments for the next five years.

Besides the uncertainty due to randomness itself, which is the canonical statistical definition, it is important to distinguish between the uncertainty due to *insufficient agreement* in the model projections, the uncertainty due to *insufficient evidence* (insufficient observational data to constrain the model projections or insufficient number of simulations to infer projections or insufficient lack of understanding of the physical processes), and the uncertainty induced by *insufficient literature*, which refers to the lack of published analyses of projections. For instance, models may agree on a projected change, but if this change is controlled by processes that are not well understood and validated in the present climate, then there is an inherent uncertainty in the projections, no matter how good the model agreement may be. Similarly, available model projections may agree in a given change, but the number of available simulations may restrain the reliability of the inferred agreement (e.g., because the analyses need to be based on daily data which may not be available from all modelling groups).

Uncertainty analysis of the CMIP3 MME in AR4 focused essentially on the seasonal mean and inter-model standard deviation values (Christensen et al., 2007; Meehl et al., 2007b; Randall et al., 2007). Where the ensemble mean projected climate change was larger than the standard deviation, the signal was generally considered to be ‘robust’. In addition, confidence was assessed in the AR4 through simple quantification of the number of models that show agreement in the sign of a specific climate change (e.g., sign of the change in frequency of extremes) – assuming that the greater the number of models in agreement, the greater the robustness. However, the ensemble was strictly an “ensemble of opportunity”, without sampling protocol and the possible dependence of different models on one another (e.g., due to shared parameterizations) was not assessed (Knutti et al., 2010a). Furthermore, this particular metric, that assesses sign agreement only, can provide misleading conclusions in cases, for example, where the projected changes are near zero. We nonetheless use a similar metric in several of the figures of the IPCC SREX (indicating “likely” changes when at least 66% of the models agree on the sign of change).

Post-AR4 studies have concentrated more on the use of the MME in order to better characterize uncertainty in climate change projections, including those of extremes (Kharin et al., 2007; Gutowski et al., 2008a; Perkins et al., 2009), and new techniques have been developed for exploiting the full ensemble information, in some cases using observational constraints to construct probability distributions (Tebaldi and Knutti, 2007; Tebaldi and Sanso, 2009). Perturbed-physics ensembles have also become available (e.g., Collins et al., 2006; Murphy et al., 2007) and used to examine projected changes in extremes and their uncertainties (Barnett et al., 2006; Clark et al., 2006; Burke and Brown, 2008; Clark et al., 2011). Advances have also been made in developing probabilistic information at regional scales from the AOGCM simulations, but there has been rather less development extending this to probabilistic downscaled regional information and to extremes (Fowler et al., 2007b; Fowler and Ekstrom, 2009). One recent example of such projections provides probability distributions of changes in various parameters including the wettest and hottest days of each season for 25 km grid squares across the UK (Murphy et al., 2009). In general, downscaling methods are maturing and being more widely applied (despite being still restricted in terms of geographical coverage, Maraun et al., 2010).

Both statistical and dynamical downscaling methods are affected by the uncertainties which affect the global models, and a further level of uncertainty associated with the downscaling step also needs to be taken into consideration (see also Sections 3.2.3.1 and 3.2.3.2). The increasing availability of coordinated RCM simulations for different regions permits more systematic exploration of dynamical downscaling uncertainty. Such simulations are available for Europe (e.g., Christensen and Christensen, 2007; van der Linden and Mitchell, 2009) and a few other regions such as North America (Mearns et al., 2009) and West Africa (van der Linden and Mitchell, 2009; Hourdin et al., 2010). RCM intercomparisons have also been undertaken for a number of regions including Asia (Fu et al., 2005), South America (Menendez et al., 2010) and the Arctic (Inoue et al., 2006). A new series of co-ordinated simulations covering the globe is planned (Giorgi et al., 2009). Increasingly, RCM output from co-ordinated simulations is made available at the daily timescale, facilitating the analysis of some extreme events. Ensuring adequate sampling of RCMs may be more important for extremes than for changes in mean values (Frei et al., 2006; Fowler et al., 2007b). Natural variability, for example, has been shown to make a significant contribution to the spectrum of variability on at least multi-annual timescales and potentially up to multi-decadal timescales in the case of European projections of precipitation extremes (Kendon et al., 2008). Some comparisons between statistical and dynamical downscaling results are available but scarce (Section 3.2.3.1).

START BOX 3.1 HERE

Box 3.1: Variations in Confidence in Projections of Climate Change: Mean vs. Extremes, Variables, Scale

Comparisons of observed and simulated climate demonstrate good agreement for some climate variables such as mean temperature, especially at large horizontal scales (e.g., Räisänen, 2007). For instance, Figure 9.12 of the AR4 (Hegerl et al., 2007) compares the ability of 14 climate models to simulate the decadal variations of mean temperature through the 20th century. When the models included both natural and anthropogenic forcings, they consistently reproduced the decadal variations in global mean temperature. Without the anthropogenic influences the models consistently failed to reproduce the decadal temperature variations. However, when the same models' abilities to simulate the temperature variations on smaller domains are assessed, although the mean temperature produced by the ensemble generally tracked the observed temperature changes, the consistency between the models was poorer than was the case for the global mean (Figure 9.12, Hegerl et al., 2007). We can conclude that the smaller the spatial domain for which simulations or projections are being prepared, the less confidence we should have in these projections.

This increased uncertainty at smaller scales results from larger internal variability at smaller scales or "noise" (i.e., natural variability unrelated to external forcings) and increased model uncertainty (i.e., less consistency between models) at these scales (Hawkins and Sutton, 2009). The latter factor is largely due to the role of unresolved processes (representations of clouds, convection, land-surface processes, see also Section 3.2.3). Hawkins and Sutton (2009) also point out regional variations in these aspects: in the tropics the temperature signal expected from anthropogenic factors is large relative to the model uncertainty and the natural variability, compared with higher latitudes. Figure 9.12 from AR4 (Hegerl et al., 2007) also shows that the models are more consistent in reproducing decadal temperature variations in the tropics than at higher latitudes, even though the magnitudes of the temperature trends are larger at higher latitudes.

Uncertainty in projections also depends on the considered variables, phenomena or impacts (Sections 3.3. to 3.5.). There is more model uncertainty for variables other than temperature, for instance precipitation (Räisänen, 2007; Hawkins and Sutton, 2011, see also Section 3.2.3). And the situation is more difficult again for extremes. For instance, climate models simulate observed changes in extreme temperatures relatively well, but the frequency, distribution and intensity of heavy precipitation is more poorly simulated (Randall et al., 2007) as are observed changes in heavy precipitation (e.g., Alexander and Arblaster, 2009). Also, projections of changes in temperature extremes tend to be more consistent across climate models than for (wet and dry) precipitation extremes (Tebaldi et al., 2006; Orlowsky and Seneviratne, 2011) and significant inconsistencies are also found for projections of agricultural (soil moisture) droughts (Wang, 2005; see also Box 3.2). For some other extremes, such as tropical cyclones, differences in the regional-scale climate change projections between models can lead to marked differences in projected tropical cyclone activity associated with anthropogenic climate change (Knutson et al., 2010), and thus decrease confidence in projections of changes in that extreme.

The relative importance of various causes of uncertainties in projections is somewhat different for earlier compared with later future periods. For some variables (mean temperature, temperature extremes), the choice of emission scenario becomes more critical than model uncertainty for the second part of the 21st century (Tebaldi et al., 2006; Hawkins and Sutton, 2009, 2011) though this does not apply for mean precipitation and some precipitation-related extremes (Tebaldi et al., 2006; Hawkins and Sutton, 2009), and has in particular not been evaluated in detail for a wide range of extremes. Users need to be aware of such issues in deciding the range of uncertainties that it is appropriate to consider for their particular risk or impacts assessment

In summary, confidence in climate change projections depends on the considered (temporal and spatial) scale, variable and whether one considers extremes or mean quantities. Confidence is highest for temperature, especially on global scales, and decreases when other variables are considered, and when we focus on smaller spatial domains (Tables 3.1 and 3.3.). Confidence in projections for extremes is weaker than for projections of long-term averages.

END BOX 3.1 HERE

[INSERT TABLE 3.3 HERE

Table 3.3: Projected regional changes in temperature and precipitation (including dryness) extremes. See Figure 3.2 for definitions of regions. Projections are for the end of the 21st century vs end of the 20th century (i.e., 1961-1990 or 1980-1999 vs 2071-2100 or 2080-2099) and for the A2/A1B emissions scenario. Codes for the source of modelling evidence: G: Based on GCM simulations. **G**: multi-GCM. R: Based on RCM simulations. **R**: multi-GCM. <u>R</u>: multi-RCM. T06 stands for Tebaldi et al. (2006), SW08 stands for Sheffield and Wood (2008b), and OS11 stands for Orlowsky and Seneviratne (2011).]

3.3. Observed and Projected Changes of Weather and Climate Extremes

3.3.1. Temperature

Temperature is associated with several types of extremes, e.g., heatwaves and cold spells, and related impacts, e.g., on human health, ecosystems, and energy consumption (Chapter 4). Temperature extremes often occur on weather timescales which require daily or higher timescale resolution data to accurately assess possible changes (Section 3.2.1). It is important to distinguish between daily mean, maximum (i.e., daytime), and minimum (nighttime) temperature, as well as between cold and warm extremes, due to their differing impacts. Spell lengths (e.g., duration of heatwaves) are relevant for a number of impacts. Note that we do not consider here changes in diurnal temperature range or frost days, which are not typical "climate extremes". There is an extensive body of literature, regarding the mechanisms of changes in temperature extremes (e.g., Christensen et al., 2007; Meehl et al., 2007b; Trenberth et al., 2007). Heatwaves are generally caused by quasi-stationary anticyclonic circulation anomalies or atmospheric blocking (Xoplaki et al., 2003; Meehl and Tebaldi, 2004; Cassou et al., 2005; Della-Marta et al., 2007b), and/or land-atmosphere feedbacks (in transitional climate regions), whereby the latter can act as an amplifying mechanism through reduction in evaporative cooling in (Section 3.1.4), but also induce enhanced persistence due to soil moisture memory (Lorenz et al., 2010). Also snow feedbacks (Section 3.1.4) and changes in aerosols (Portmann et al., 2009) are relevant for temperature extremes. In the context of global warming enhanced temperatures, including temperature extremes, are induced by enhanced greenhouse forcing, also independently of changes in circulation patterns or surface feedbacks.

Regional historical or paleoclimatic temperature reconstructions can help place the recent instrumentally observed temperature extremes in the context of a much longer period, but literature on this topic is very sparse. For example Dobrovolny et al. (2010) reconstructed monthly and seasonal temperature over central Europe back to 1500 using a variety of temperature proxy records. They concluded that only two recent temperature extremes, the summer 2003 heatwave and the July 2006 heatwave exceed the +2 standard deviation (associated with the reconstruction method) of previous monthly temperature extremes since 1500. The coldest periods within the last five centuries occurred in the winter and spring of 1690. Another 500-year temperature reconstruction was recently completed for the Mediterranean basin by means of documentary data and instrumental observations (Camuffo et al., 2010). It suggests strong natural variability in the basin, possibly exceeding the recent warming, although discontinuities in the records limits the interpretation of this finding.

The IPCC AR4 (Trenberth et al., 2007) reported based on the CRU/UKMO datatset (Brohan et al., 2006) that global mean surface temperatures rose by 0.74°C ±0.18°C over the 100-year period 1906–2005, with a rate of warming over the 50-year period 1956–2005 almost double that over the last 100 years (0.13°C± 0.03°C vs. 0.07°C ± 0.02°C per decade). It further reported that trends were found to be stronger over land than over the oceans, and that for the globe as a whole, surface air temperatures over land rose at about double the ocean rate after 1979, with the greatest warming during winter (December to February) and spring (March to May) in the Northern Hemisphere. The AR4 also noted that the changes have not been linear and can be characterized as level prior to about 1915, a warming to about 1945, leveling out or even a slight decrease until the 1970s, and a fairly linear upward trend since then (Trenberth et al., 2007). It has been suggested that the partial levelling out and/or decrease in some regions from ca. 1945 until the end of the 1970s (and in some regions until the mid-1980s) is due to a so called "dimming" of incoming shortwave radiation in several regions, followed upon by a "brightening" phase, both linked with changes in aerosol concentrations and/or cloud cover (Pinker et al., 2005; Wild et al., 2005; Wild, 2009).

Consistent with this warming in mean temperatures, the AR4 (Trenberth et al., 2007, based on Alexander et al., 2006) reported a statistically significant increase in the numbers of warm nights and a statistically significant reduction in the numbers of cold nights for 70-75% of the land regions with data (for the spatial coverage of this dataset and the definition of warm/cold days and nights, see Section 3.2.1). Changes in the numbers of warm days and cold days also showed warming, but less marked than for nights, with ca. 40-50% of the area with data showing statistically significant changes consistent with warming (Alexander et al., 2006). Less than 1% of the area with data showed statistically significant trends in cold/warm days and nights that were consistent with cooling (Alexander et al., 2006). Trenberth et al. (2007) also reported, based on Vose et al. (2005), that from 1950 to 2004, the annual trends in minimum and maximum land-surface air temperature averaged over regions with data were 0.20°C per decade and 0.14°C per decade, respectively, and that for 1979 to 2004, the corresponding linear trends for the land areas with data were 0.29°C per decade for both maximum and minimum temperature. Based on this evidence, the IPCC AR4 (IPCC, 2007b) assessed that it was *very likely* that there had been trends towards warmer and more frequent warm days and warm nights, and towards warmer and less frequent cold days and cold nights in most land areas.

Regions which were found to depart from this overall behaviour towards more warm extremes and less cold extremes in Alexander et al. (2006) were mostly Central North America, the Eastern U.S., Southern Greenland (increase in cold days and decreases in warm days), and the southern half of South America (decrease in warm days; no data available for northern half of continent). In Central North America and the Eastern U.S. this partial tendency for a cooling trend in extremes is also consistent with a reported mean negative trend in temperatures, mostly in the spring to summer season (also termed "warming hole", e.g., Pan et al., 2004; Portmann et al., 2009). Several explanations have been suggested for this behaviour, which seems partly associated with a change in the hydrological cycle, possibly linked to soil moisture and/or aerosol feedbacks (Pan et al., 2004; Portmann et al., 2009).

More recent analyses available since the AR4 include a global study (for annual extremes) by Brown et al. (2008) based on the dataset from Caesar et al. (2006), and regional studies for North America (Peterson et al., 2008a; Meehl et al., 2009b), Central-Western Europe (since 1880; Della-Marta et al., 2007a), central and eastern Europe (Bartholy and Pongracz, 2007; Kurbis et al., 2009), the eastern Mediterranean including Turkey (Kuglitsch et al., 2010), western Central Africa, Guinea Conakry and Zimbabwe (Aguilar et al., 2009), the Tibetan Plateau (You et al., 2008) and China (You et al., 2011), Uruguay (Rusticucci and Renom, 2008), and Australia (Alexander and Arblaster, 2009). Further references can also be found in Table 3.2. Overall, these studies are consistent with the assessment of an increase in unusually warm nights and days and a reduction in unusually cold nights and days on the global basis, although they do not necessarily consider trends in all four variables, and a few single studies present statistically not significant or opposite trends to the global tendencies in some extremes, subregions, seasons, or decades. For instance, (Rusticucci and Renom, 2008) found in Uruguay a reduction of cold nights, a positive but a statistically not significant trend in warm nights, statistically not significant decreases in cold days at most investigated stations, and inconsistent trends in warm days. Together with the previous results from Alexander et al. (2006) for southern South America (see above) and further regional studies (Table 3.2), this suggests a less consistent warming tendency in South America compared to other continents. Another notable feature is that studies for Central and Southern Eastern Europe display a marked change point in trends in temperature extremes at the end of the 1970s / beginning of 1980s (Table 3.2), which for some extremes can lead to very small and/or statistically not significant overall trends since the 1960s (e.g., Bartholy and Pongracz, 2007). The timing of the change point is consistent with evidence from global trends in mean temperature (above) and may be linked to changes in incoming radiation (dimming/brightening, see above).

There are fewer studies available investigating changes in heatwave characteristics, rather than intensity or frequency of warm days or nights. Alexander et al. (2006) provided an analysis of trends in warm spells mostly in the mid- and high latitudes of the northern hemisphere. The analysis display a tendency towards longer warm spells in much of the region, with the exception of the Southeastern U.S. and Eastern Canada. Regional studies on trends in heatwaves are listed in Table 3.2. Kunkel et al. (2008) found that the U.S. has experienced a strong increase in heatwaves since 1960, although the heatwaves of the 1930s associated with extreme drought conditions still dominate the 1895-2005 time series. Kuglitsch et al. (2009) reported an increase in heatwave intensity, number and length in summer over the 1960-2006 time period in the Eastern Mediterranean. Ding et al. (2010) reported increasing numbers of heatwaves over most of China for the 1961-2007 period. The record-breaking heatwave over western and central Europe in the summer of 2003 is an example of an exceptional recent extreme (Beniston, 2004; Schär and Jendritzky, 2004). That summer (June to August) was the hottest since comparable instrumental records began around 1780 (1.4°C above the previous warmest in 1807) and perhaps the hottest since at least 1500 (Luterbacher et al., 2004). Other examples of recent extreme heatwaves include the 2006 heatwave in Europe (Rebetez et al., 2008), the 2007 heatwave in Southeastern Europe (Founda and Giannakopoulos, 2009), the 2009 heatwave in southeastern Australia (National Climate Centre), and the 2010 heatwave in Russia. Both the 2003 European heatwave (Andersen et al., 2005; Ciais et al., 2005) and the 2009 southeastern Australian heatwave were also associated with drought conditions, which can strongly enhance temperature extremes during heatwaves (see also Section 3.1.4).

Some recent analyses have led to some revisions of previously reported trends. For instance, Della-Marta et al. (2007a) found that mean summer maximum temperature change over Europe was +1.6±0.4°C, a somewhat stronger increase than reported in earlier studies. Kuglitsch et al. (2009; 2010) homogenised and analysed over 250 daily maximum and minimum temperature series in the Mediterranean region since 1960, and found that after homogenisation the positive trends in the frequency of hot days and heatwaves in the Eastern Mediterranean were higher than reported in earlier studies. This was due to the correction of many warm biased temperature records in the region during the 1960s and 1970s.

In summary, regional and global analyses of temperature extremes on land generally show recent changes consistent with a warming climate on the global scale, in agreement with the previous assessment from AR4. Only a few regions show changes in temperature extremes consistent with cooling, most notably for some extremes in Central North America, the Eastern U.S., and also parts of South America. Based on the available evidence we can state that it is *very likely* that there has been an overall decrease in the number of unusually cold days and nights and *very likely* that there has been an overall increase in the number of unusually warm days and nights *on the global scale*, i.e., for land areas with data (corresponding to ca. 70-80% of all land areas, see Table 3.2). It is *likely* that this statement applies at the continental scale in North America and Europe, and *very likely* that it applies in Australia (Table 3.2). However, some subregions on these continents have had warming trends in temperature extremes that were small or not statistically significant (e.g., Southeastern Europe), and a few subregions have had cooling trends in some temperature extremes (e.g., Central North America and Eastern U.S.). Asia also shows trends consistent with warming in most of the continent, but which are assessed here to be of *medium confidence* because of lack of literature for several regions beside the global study from Alexander et al. (2006). Most of Africa is insufficiently well sampled to allow an overall likelihood statement to be made at the continental scale, although most of the regions on this continent for which data are available have exhibited warming in temperature extremes (Table 3.2). In South America, both lack of data, and partial inconsistencies in the reported trends imply *low confidence* in the overall trends at the continental scale (Table 3.2). Furthermore, based on a limited number of regional analyses and implicit from the documented changes in daily temperatures, it appears that warm spells, including heatwaves defined in various ways, have *likely* increased in frequency since the middle of the 20th century in many regions with some exceptions (Table 3.2).

The AR4 (Hegerl et al., 2007) concluded that surface temperature extremes have *likely* been affected by anthropogenic forcing. This assessment was based on multiple lines of evidence of temperature extremes at the global scale including the reported increase in the number of warm extremes and decrease in the number of cold extremes on that scale (Alexander et al., 2006). There was also evidence that anthropogenic forcing may have statistically significantly increased the likelihood of extreme temperatures (Christidis et al., 2005) and of the 2003 European heat wave (Stott et al., 2004).

Recent studies on attributions of changes in temperature extremes have tended to reaffirm the conclusions reached in the AR4. Alexander and Arblaster (2009) found that trends in 'warm nights' over Australia could only be reproduced by a coupled model that included anthropogenic forcings. Meehl et al. (2007b) showed that most of the observed changes in temperature extremes for the second half of the 20th century over the U.S. can be attributed to human activity. They compared observed changes in the number of frost days, the length of growing season, the number of warm nights, and the heatwave intensity with those simulated in a nine member multi-model ensemble simulation. The decrease of frost days, an increase in growing season length, and an increase in heatwave intensity all show similar changes over the U.S. in 20th century experiments that combine anthropogenic and natural forcings, though the relative contributions of each are unclear. Results from two global coupled climate models with separate anthropogenic and natural forcing runs indicate that the observed changes are simulated with anthropogenic forcings, but not with natural forcings (even though there are some differences in the details of the forcings). Zwiers et al. (2011) compared observed annual temperature extremes including annual maximum daily maximum and minimum temperatures, and annual minimum daily maximum and minimum temperatures with those simulated responses to anthropogenic (ANT) forcing or anthropogenic and natural external forcings combined (ALL) by multiple GCMs. They fitted probability distributions (Section 3.1.2) to the observed extreme temperatures with a time-evolving pattern of location parameters as obtained from the model simulation, and found that both anthropogenic influence and combined influence of anthropogenic and natural forcing can be detected in all four extreme temperature variables at the global scale over the land, and also regionally over many large land areas. They concluded that the influence of anthropogenic forcing has had a detectable influence on extreme temperatures at global and regional scales. Globally, waiting times for events that were expected to recur once every 20 years in the 1960s are now estimated to exceed 30 years for extreme annual minimum daily maximum temperature and 35 years for extreme annual minimum daily minimum temperature, and to have decreased to less than 10 or 15 years for annual maximum daily minimum and daily maximum temperatures respectively (Figure 3.3). However, the available detection and attribution studies for extreme maximum temperatures (Christidis et al., 2011; Zwiers et al., 2011) suggest that the models over-estimate changes in these extremes during the late 20th century.

[INSERT FIGURE 3.3 HERE
Figure 3.3: Estimated waiting time (years) and their 5% and 95% uncertainty limits for 1960s 20-yr return values of annual extreme daily temperatures in the 1990s climate (see text for more details). From Zwiers et al. (2011). Red, green, blue, pink error bars are for annual minimum daily minimum temperature (TNn), annual maximum daily minimum temperature (TNx), annual minimum daily maximum temperature (TXn), and annual maximum daily maximum temperature (TXx), respectively. Grey areas indicate insufficient data.]

Regarding projections of extreme temperatures, the AR4 (Meehl et al., 2007b) noted that cold episodes were projected to decrease significantly in a future warmer climate and considered *very likely* that heatwaves would be more intense, more frequent and last longer in a future warmer climate. Post-AR4 studies of temperature extremes have utilised larger model ensembles (Kharin et al., 2007; Sterl et al., 2008; Orlowsky and Seneviratne, 2011) and generally confirm the conclusions of AR4, while also providing more specific assessments both in terms of the range of considered extremes and the level of regional detail (see also Table 3.3).

There are few *global* analyses of multi-model projections in temperature extremes available in the literature. The study by Tebaldi et al. (2006), which was referenced in the AR4 (Figures. 10.18 and 10.19 in Meehl et al., 2007b), provided global analyses of projected changes (A1B scenario) in several extremes based on 9 GCMs. For temperature extremes, analyses were provided for heatwave lengths (using the HWDImax index, see Section 3.1.2 and discussion) and warm nights. Stippling was provided when 5 out of 9 models displayed statistically significant changes of the same sign. Orlowsky and Seneviratne (2011) recently updated the analysis from Tebaldi et al. (2006) for the full ensemble of GCMs (23 in total) that contributed A2 scenarios to the CMIP3, using a larger number of extreme indices (for temperature; HWDImean and WSDI in addition to HWDImax, see Section 3.1.2; and percentages of warm days, cold days, warm nights, cold nights, days with temperature greater than 30°C and nights with temperatures greater than 20°C), using other thresholds for display and stippling of the figures (no results displayed if less than 66% of the models agree on the sign of change, stippling provided only for 90% model agreement), and providing seasonal analyses. This analysis confirmed that strong agreement exists between the various GCM projections for temperature-related extremes, with projected increases of warm day occurrences (Figure 3.4) and heatwave length (see discussion in Section 3.1.2), and decreases of cold extremes (Figure 3.5). Moderate temperature extremes on land were found to be projected to warm faster than global annual mean temperature in many regions and seasons (Figure 3.1), implying large changes in extremes in some places, even for a global warming of 2°C or 3°C (with scaling factors for the SRES A2 scenario ranging between 0.5 and 2.5 for moderate seasonal extremes, Figure 3.1). Based on the analyses of Tebaldi et al. (2006) and Orlowsky and Seneviratne (2011), as well as physical considerations, we assess that increases in the number of hot extremes and decreases in the number of cold extremes (defined with respect to present regional climate, i.e., 1961-1990 reference period) are *virtually certain* on the global scale. Further, given the assessed changes in hot and cold extremes and available analyses of projected changes in heatwave length in the two studies, we assess that it is *very likely* that the length, frequency and/or intensity of heatwaves will increase on the global scale.

Another global study of changes in extremes based on the CMIP3-ensemble is provided in Kharin et al. (2007), which focuses on changes in annual extremes (20-year extreme values) based on 14 GCMs for temperature extremes and 16 GCMs for precipitation extremes and employed the SRES A2, A1B, and B1 emissions scenarios. This analysis projects increases in the temperature of the one-in-twenty years annual extreme hottest day of 2-6°C (depending on region; Figure 3.6 adapted from Kharin et al., 2007) , and strong reductions in the waiting times of this extreme event. However, as noted above, the limited number of relevant detection and attribution studies suggest that models may over-estimate the changes in temperature extremes, and our assessments take this into account by weakening the uncertainty assessments from what would be derived by uncritical acceptance of the projections in Figure 3.6. The uncertainty estimates are also reduced to reflect the possibility that some important processes relevant to extremes may be missing or be poorly represented in models, as well as the fact that the model projections considered in this study did not correspond to the full CMIP3 ensemble. Globally we assess that under the A2 and A1B scenarios a one-in 20 year annual extreme hot day is *likely* to become a one-in-two year annual extreme by the end of the 21st century in most regions, except in the high latitudes of the northern hemisphere where it is *likely* to become a one-in-five year annual extreme (Figure 3.6b, newly computed based on material from Kharin et al., 2007). Further, we assess that under the more moderate B1 scenario a current one-in-20 year extreme would *likely* become a one-in-five year event (and a one-in-ten year event in northern hemisphere high latitudes).

In the following paragraph, regional assessments of projected changes in temperature extremes are provided. More details are found in Table 3.3. For North America, the U.S. Climate Change Science Program (CCSP) reached the following conclusions (using IPCC likelihood terminology) regarding projected changes in temperature extremes by the end of the 21st century (Gutowski et al., 2008a): 1) Abnormally hot days and nights and heat waves are *very likely* to become more frequent; 2) Cold days and cold nights are *very likely* to become much less frequent; 3) For a mid-range scenario of future greenhouse gas emissions, a day so hot that it is currently experienced only once every 20 years would occur every three years by the middle of the century over much of the continental U.S. and every five years over most of Canada; by the end of the century, it would occur every other year or more. For Australia, the CMIP-3 ensemble was projected increases in warm nights (15–40% by the end of the 21st century) and heat wave duration, together with a decrease in the number of frost days (Alexander and Arblaster, 2009). Inland regions show greater warming compared with coastal zones (Suppiah et al., 2007; Alexander and Arblaster, 2009) and large increases in the number of days above 35°C or 40°C are indicated (Suppiah et al., 2007). For the entire South American region, a study with a single RCM projected more frequent warm nights and fewer cold nights (Marengo et al., 2009a). Several studies of regional and global projections of changes in extremes are available for the European continent (see also Table 3.3.). Analyses of both global and regional model outputs show major increases in warm temperature extremes across the Mediterranean including events such as hot days (Tmax >30°C) and tropical nights (Tmax >20°C) (Giannakopoulos et al., 2009; Tolika et al., 2009). Comparison of RCM projections with data for 2007 (the hottest summer in Greece in the instrumental record with a record daily Tmax observed value of 44.8°C) indicates that the distribution for 2007 lies entirely within the distribution for 2071–2100 - thus 2007 might be considered a ‘normal’ summer of the future (Founda and Giannakopoulos, 2009; Tolika et al., 2009). Beniston et al. (2007) concluded from an analysis of RCM output that regions such as France and Hungary, may experience as many days per year above 30°C as currently experienced in Spain and Sicily. In this RCM ensemble, France was the area with the largest projected warming in the uppermost percentiles of daily summer temperatures although the mean warming is greatest in the Mediterranean (Fischer and Schär, 2009). New results from an RCM ensemble project increases in the amplitude, frequency and duration of health-impacting heatwaves, especially in southern Europe (Fischer and Schär, 2010). Overall these regional assessments are consistent with the global assessments provided above.

[INSERT FIGURE 3.4 HERE

Figure 3.4: Projected annual and seasonal changes of three indices for Tmax: Fraction of warm days, fraction of cold days, and fraction of days with Tmax > 30°C; CMIP3 projections, 2080-2100 time frame minus 1980-2000 time frame

(projections for A2 scenario, relative to late 20th century (20C3M) simulations), annual (top), DJF (middle) and JJA (bottom). Shading is only applied for areas where at least 66% of the models agree in the sign of the change; stippling is applied for regions where at least 90% of all models agree in the sign of the change [from Orlowsky and Seneviratne, 2011, after Tebaldi et al., 2006].

[INSERT FIGURE 3.5 HERE
Figure 3.5: Projected annual and seasonal changes of three indices for Tmin: Fraction of warm nights, fraction of cold nights, and fraction of days with Tmin > 20°C; CMIP3 simulations, 2080-2100 time frame minus 1980-2000 time frame (projections for A2 scenario, relative to late 20th century (20C3M) simulations), annual (top), DJF (middle) and JJA (bottom). Shading is only applied for areas where at least 66% of the models agree in the sign of the change; stippling is applied for regions where at least 90% of all models agree in the sign of the change [from Orlowsky and Seneviratne, 2011, after Tebaldi et al., 2006].

[INSERT FIGURE 3.6 HERE
Figure 3.6: (a, top) Projected changes from the late-twentieth-century 20-year return values of annual maximum of the daily maximum temperature in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the CMIP3, under three different SRES emission scenarios B1 (blue), A1B (green) and A2 (red); units in °C. Adapted from the analysis in Kharin et al. (2007). The vertical extent of the whiskers shows the range of projected changes from all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Model projections suggest that the the 20-year extreme annual daily maximum temperature will increase by about 2°C by mid-21st century and by about 4°C by late-21st century, depending on the region.
(b, bottom) Projected waiting times for late-twentieth-century 20-year return values of annual maximum of the daily maximum temperature in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the CMIP3, under three different SRES emission scenarios B1, A1B and A2 Adapted from the analysis in Kharin et al. (2007). The vertical extent of the whiskers shows the range of projected changes from all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Model projections suggest that the waiting time for a late 20th century 20-year extreme annual daily maximum temperature will be reduced to about 2-20 years by mid-21st century and by about 1-5 years by late-21st century, depending on the region. Two global domains for which projections are shown are: the entire globe including the oceans, and the global land areas.]

Temperature extremes were the type of extremes projected to change with most confidence in the AR4 (IPCC, 2007a). This is confirmed regarding the sign of change with more recent analyses (Figure 3.4 and 3.5, from Orlowsky and Seneviratne, 2011), although there is a large spread with respect to the magnitude of changes both due to emission scenario and climate model uncertainty (Figures 3.6a and 3.6b, adapted from Kharin et al., 2007). If changes in temperature extremes scale with changes in mean temperature (i.e., simple shifts of the probability distribution), we can infer that it is *virtually certain* that hot (cold) extremes will increase (decrease) in the coming decades (if these extremes are defined with respect to the 1961-1990 climate). Changes in the tails of the temperature distributions may not scale with changes in the mean in some regions (Section 3.1.6), though in most such reported cases hot (cold) extremes tend to increase (decrease) more than mean temperature, and thus the above statement for extremes (*virtually certain* increase in hot extremes and decrease in cold extremes) still applies. Central and Eastern Europe is a region where evidence suggests that projected changes in temperature extremes result from both changes in the mean as well as from changes in the shape of the probability distributions (Schär et al., 2004). The main mechanism for the widening of the distribution is linked to the drying of the soil in this region (Sections 3.1.4 and 3.1.6). Furthermore, remote surface heating may induce circulation changes that modify the temperature distribution (Haarsma et al., 2009). Other local, mesoscale and regional feedback mechanisms, in particular with land surface conditions (beside soil moisture, also with vegetation and snow; Section 3.1.4) and aerosol concentrations (Ruckstuhl and Norris, 2009) may enhance the uncertainties in temperature projections. Some of these processes occur on a small scale un-resolved by the models (Section 3.2.3). In addition, lack of observational data (e.g., for soil moisture and snow cover, see Section 3.2.1) reduces the possibilities to validate climate models (e.g., Roesch, 2006; Boe and Terray, 2008; Hall et al., 2008; Brown and Mote, 2009). Regarding mesoscale processes, lack of information may also affect confidence in projections. One example is changes in Mediterranean heatwaves which are suggested to have the largest impact in coastal areas, due to the role of enhanced relative humidity for health impacts (Diffenbaugh et al., 2007; Fischer and Schär, 2010). But it is not clear how this pattern may or may not be moderated by sea breezes (Diffenbaugh et al., 2007).

In summary, since 1950 it is *very likely* that there has been an overall decrease in the number of unusually cold days and nights and an overall increase in the number of unusually warm days and nights on the global scale, i.e., for land areas with data. It is *likely* that such changes have also occurred at the continental scale in North America and Europe, and *very likely* in Australia. There is *medium confidence* of a warming trend in temperature extremes in much of Asia. There is *low confidence* in trends for Africa, because of lack of data and

studies. In South America, lack of data, and some inconsistencies between reported trends imply *low confidence* in the trends at the continental scale. It is *likely* that the number of warm spells, including heatwaves, have increased since the middle of the 20th century in many (but not all) regions. On the global scale, climate change projections indicate a *virtually certain* increase in the number of hot extremes and *virtually certain* decrease in the number of cold extremes (defined with respect to present regional climate, i.e., 1961-1990 reference period). This is mostly linked with mean changes in temperatures, although changes in temperature variability can play an important role in some regions. It is *very likely* that the length, frequency and/or intensity of heatwaves (defined with respect to present regional climate) will continue to increase on the global scale. Projected changes on sub-continental scales are less certain than is the case for the global scale. A one-in-20 year annual hottest day is *likely* to become a one-in-two year annual extreme by the end of the 21st century in most regions, except in the high latitudes of the northern hemisphere where it is *likely* to become a one-in-five year annual extreme. Moderate temperature extremes on land are projected to warm faster than global annual mean temperature in many regions and seasons. A mean global warming of 2°C or 3°C is *likely* to lead to much larger increases in some temperature extremes in certain regions and seasons (with scaling factors for the SRES A2 scenario *likely* ranging between 0.5 and 2.5 for moderate seasonal extremes). Nevertheless, mean global warming does not necessarily imply warming in all regions and seasons.

3.3.2. Precipitation

This section addresses changes in short-term extreme or heavy precipitation events. Changes in mean (or total) precipitation that can lead to drought (i.e., associated with lack of precipitation) are considered in Section 3.5.1. Because climates are so diverse across different parts of the world, it is difficult to provide a single definition of extreme or heavy precipitation. In general, two different approaches have been used: 1) relative thresholds such as percentiles and return values (typically the 95th percentile) and 2) absolute thresholds (e.g., 50.8 mm (2 inches)/day of rain in the U.S., and 50mm/day or 100mm/day of rain in China). For more details on the respective drawbacks and advantages of these two approaches, see Section 3.1. Note that we do not distinguish between rain and snowfall (both considered as contributors to overall extreme precipitation events), but do distinguish changes in hail from other precipitation types. Increases in public awareness and changes in reporting practices have led to inconsistencies in the record of severe thunderstorms and hail that make it difficult to detect trends in the intensity or frequency of these events (Kunkel et al., 2008). Furthermore, weather events such as hail are not well captured by current monitoring systems and, in some parts of the world, the monitoring network is very sparse (Section 3.2.1), resulting in considerable uncertainty in the estimates of extreme precipitation. There are also known biases in precipitation measurements, mostly leading to rain undercatch.

Little evidence of paleo and historical changes in heavy precipitation is available to place recent variations into context. An overview of mid- to late-Holocene climate change (Wanner et al., 2008) suggested a pronounced weakening of the monsoon systems in Africa and Asia and increasing dryness and desertification on both continents, which accompanied a progressive southward shift of the Northern Hemisphere summer position of the Intertropical Convergence Zone. A study for Europe (Pauling and Paeth, 2007) suggested that there were large fluctuations in wet winters over the last 300 years, and that 1951-2000 displays more extreme wet winters than other 50-year periods in the 300 preceding years with the exception of 1701-1750. A study for the Middle East (Black et al., 2010) reported decreased winter rainfall in southern Europe and the Middle East and increased rainfall further north during the Holocene, caused by a poleward shift of the North Atlantic storm track and a weakening of the Mediterranean storm track. A study for southern Spain (Rodrigo et al., 1999) suggested that the wettest periods occurred at the end of 16th century, the beginning of 17th century, and at the end of 19th century, while the driest periods in the pre-instrumental era occurred during the first half of the 16th century, and around 1750.

The AR4 (Trenberth et al., 2007) concluded that it was *likely* that there had been increases in the number of heavy precipitation events (e.g., 95th percentile) within many land regions, even in those where there had been a reduction in total precipitation amount, consistent with a warming climate and observed significant increasing amounts of water vapour in the atmosphere. Increases had also been reported for rarer precipitation events (1 in 50 year return period), but only a few regions had sufficient data to assess such trends reliably. However, the AR4 (Trenberth et al., 2007) also stated that "Many analyses indicate that the evolution of rainfall statistics through the second half of the 20th century is dominated by variations on the interannual to inter-decadal time scale and that trend estimates are spatially incoherent (Manton et al., 2001; Peterson et al., 2002; Griffiths et al., 2003; Herath and Ratnayake, 2004)". Overall, as highlighted in Alexander et al. (2006), the observed changes in precipitation extremes were found at the time to be much less spatially coherent and statistically significant compared to observed changes in temperature extremes: Although statistically significant trends towards stronger precipitation extremes were generally found for a larger fraction of the land area than trends towards weaker precipitation extremes, statistically significant changes in precipitation indices for the overall land areas with data were only found for precipitation intensity, and not for other considered indices (Alexander et al., 2006).

Post-AR4 studies updated the results assessed at that time, with more regions being covered (Table 3.2). Overall, this additional evidence confirms that more locations and studies show an increase than a decrease in extreme precipitation, but that there are also wide regional and seasonal variations, and that trends in many regions are not statistically significant (Table 3.2). More detailed regional assessments are provided hereafter.

Recent studies on past and current changes of precipitation extremes in North America, some of which are included in the recent assessment of the U.S. Climate Change Science Program (CCSP) report (Kunkel et al., 2008), have reported an increasing trend over the last half century. Based on station data from Canada, the U.S., and Mexico, Peterson et al. (2008b) reported that heavy precipitation has been increasing over 1950–2004, as well as the average amount of precipitation falling on days with precipitation. For the contiguous U.S., DeGaetano (2009) showed a 20% reduction in the return period for extreme precipitation of different return levels over 1950–2007; Gleason et al. (2008) reported an increasing trend in the area experiencing a much above-normal proportion of heavy daily precipitation from 1950 to 2006; Pryor et al. (2009) provided evidence of increases in the intensity of events above the 95th percentile during the 20th century, with a larger magnitude of the increase at the end of the century. The largest trends towards increased annual total precipitation, number of rainy days and intense precipitation (e.g., fraction of precipitation derived from events in excess of the 90th percentile value) were focused on the central plains/northwestern Midwest (Pryor et al., 2009). In the core of the North American monsoon region in northwest Mexico, statistically significant positive trends were found in daily precipitation intensity and seasonal contribution of daily precipitation greater than its 95th percentile in the mountain sites for the period 1961–1998. However, no statistically significant changes were found in coastal stations (Cavazos et al., 2008). Overall, the evidence indicates a *likely* increase in observed heavy precipitation in many regions in North America, despite statistically non-significant trends and some decreases in some subregions (Table 3.2).

There is overall *low confidence* in trends for the whole of Central and South America (Table 3.2). Positive trends in extreme rainfall events are evident in parts of southern South America (Dufek and Ambrizzi, 2008; Marengo et al., 2009b; Re and Ricardo Barros, 2009; Sugahara et al., 2009). But negative trends have been observed in winter extreme precipitation in some regions (Penalba and Robeldo, 2010).

There is *medium confidence* in trends in heavy precipitation in Europe, due to partly inconsistent signals across studies and regions, especially in summer (Table 3.2). Winter extreme precipitation has increased in part of the continent, in particular in Central-Western Europe and European Russia (Zolina et al., 2009), but the trend in summer precipitation has been weak or not spatially coherent (Moberg et al., 2006; Bartholy and Pongracz, 2007; Maraun et al., 2008; Pavan et al., 2008; Zolina et al., 2008; Costa and Soares, 2009; Kysely, 2009; Durão et al., 2010; Rodda et al., 2010). Increasing trends in 90th, 95th and 98th percentiles of daily winter precipitation over 1901–2000 were found (Moberg et al., 2006), which has been confirmed by more detailed country-based studies for the United Kingdom (Maraun et al., 2008), Germany (Zolina et al., 2008), Belgium (Ntegeka and Willems, 2008), Central and Eastern Europe (Bartholy and Pongracz, 2007; Kysely, 2009), while decreasing trends have been found in some regions such as northern Italy (Pavan et al., 2008), Poland (Lupikasza, 2010) and some Mediterranean coastal sites (Toreti et al., 2010). Uncertainties are overall larger in Southern Europe and the Mediterranean, where there is *low confidence* in the trends (Table 3.2). A recent study (Zolina et al., 2010) has indicated that there has been an increase by about 15-20% in the persistence of wet spells over most of Europe over the last 60 years, which was not caused by an increase of the total number of wet days.

There is overall *low confidence* in trends in heavy precipitation in Asia, both on the continental and regional scale for most regions (Table 3.2; see also Alexander et al., 2006). In the Asia-Pacific region, no systematic spatially coherent trends in the frequency and duration of extreme precipitation events have been found (Choi et al., 2009). However, statistically significant positive and negative trends were observed at sub-regional scales within this region. Heavy precipitation increased in Japan during 1901-2004 (Fujibe et al., 2006), and in India (Rajeevan et al., 2008; Krishnamurthy et al., 2009) especially during the monsoon seasons (Sen Roy, 2009; Pattanaik and Rajeevan, 2010). Both statistically significant increases and decreases in extreme precipitation have been found in China over the period 1951-2000 (Zhai et al., 2005) and 1978-2002 (Yao et al., 2008). Heavy precipitation increased over the southern and northern Tibetan Plateau but decreased in the central Tibetan Plateau during 1961–2005 (You et al., 2008). No spatially coherent trends in extreme precipitation during 1950-2003 over mid-East countries were found (Zhang et al., 2006). In Peninsular Malaysia during 1971–2005 the intensity of extreme precipitation increased and frequency decreased, while the trend in the proportion of extreme rainfall over total precipitation was not statistically significant (Zin et al., 2009).

In Southern Australia, there has been a *likely* increase in heavy precipitation in many areas, except where mean precipitation has decreased (Table 3.2), but there is *low confidence* in the trends in Northern Australia due to lack of literature (Table 3.2). Extreme summer rainfall over the northwest of the Swan-Avon River basin in western Australia increased over 1950-2003 while extreme winter rainfall over the southwest of the basin decreased (Aryal et al., 2009).

There is *low to medium confidence* in regional trends in heavy precipitation in Africa due to partial lack of literature and data, and due to lack of consistency in reported patterns in some regions (Table 3.2). The IPCC AR4 (Trenberth et al.,

2007) reported an increase in heavy precipitation over southern Africa, but this appears to depend on region and precipitation index examined (Kruger, 2006; New et al., 2006; Seleshi and Camberlin, 2006; Aguilar et al., 2009; Camberlin et al., 2009). Central Africa exhibited a decrease in heavy precipitation over the last half century (Aguilar et al., 2009), however data coverage for large parts of the region was poor. The IPCC AR4 reported a decrease in heavy precipitation in east Africa, but Camberlin et al., (2009) found no trend in daily precipitation intensity. There were decreasing trends in heavy precipitation over parts of Ethiopia over the period 1965-2002 (Seleshi and Camberlin, 2006).

Changes in hail occurrence are generally difficult to quantify because hail occurrence is not well captured by the monitoring system. Sometimes, changes in the environment conditions conducive to hail occurrence are used to infer changes in hail occurrence. However, the atmospheric conditions are typically estimated from reanalyses or from radiosonde data that are associated with high uncertainty. As a result, assessment of changes in hail frequency is difficult. Over the United States, DeRubertis (2006) found widespread trends toward enhanced atmospheric instability in summer and Changnon and Changnon (2000) found five types of temporal variations of hail frequency. For severe thunderstorms in the region east of the Rocky Mountains in the United States, Brooks and Dotzek (2008) found strong variability but no clear trend in the past 50 years. Cao (2008) identified a robust upward trend in hail frequency over Ontario, Canada. Kunz et al. (2009) found that both hail damage days and convective instability increased during 1974-2003 in a state in southwest Germany. Piani et al. (2010) identified an increasing trend in hailstorm frequency in Italy during 1961-2003. Xie et al. (2008) identified no trend in the mean annual hail days in China from 1960 to early 1980s but a statistically significant decreasing trend afterwards.

In summary, it is *likely* that there has been statistically significant increases in the number of heavy precipitation events (e.g., 95th percentile) in more regions than there has been statistically significant decreases, but there are strong regional and subregional variations in the trends. In particular, many regions present statistically non-significant or negative trends, and there are also variations between seasons (more consistent trends in winter than in summer in Europe). The overall most consistent trends towards heavier precipitation events are found in North America (*likely* increase over the whole continent). This overall assessment is consistent with that of the AR4.

The observed changes in heavy precipitation appear to be consistent with the expected response to anthropogenic forcing (increase due to enhanced moisture content in the atmosphere) but a direct cause-and-effect relationship between changes in external forcing and extreme precipitation had not been established at the time of the AR4. As a result, the AR4 concluded only that it is *more likely than not* that anthropogenic influence had contributed to a global trend towards increases in the frequency of heavy precipitation events over the second half of the 20th century (Hegerl et al., 2007).

New research since the AR4 provides more evidence of anthropogenic influence on various aspects of the global hydrological cycle (Stott et al., 2010; see also Section 3.2.2.2), which is directly relevant to extreme precipitation changes. In particular, an anthropogenic influence on atmospheric moisture content is detectable (Santer et al., 2007; Willett et al., 2007; see also Section 3.2.2.2). Wang and Zhang (2008) show that winter season maximum daily precipitation in North America appears to be statistically significantly influenced by atmospheric moisture content, with an increase in moisture corresponding to an increase in maximum daily precipitation. This behaviour has also been seen in model projections of extreme winter precipitation under global warming (Gutowski et al., 2008b). The thermodynamic constraint based on the Clausius-Clapeyron relation is a good predictor for extreme precipitation changes in a warmer world in regions where the nature of the ambient flows change little (Pall et al., 2007). This may support the judgment that the observed increase in extreme precipitation may, in part, be attributable to anthropogenic influence. However, the thermodynamic constraint may not be a good predictor in regions with circulation changes such as mid- to higher-latitudes (Meehl et al., 2005) and the tropics (Emori and Brown, 2005). Additionally, changes of precipitation extremes with temperature also depend on changes in the moist-adiabatic temperature lapse rate, in the upward velocity, and in the temperature when precipitation extremes occur (O'Gorman and Schneider, 2009a, b; Sugiyama et al., 2010). This may explain why there have not been increases in precipitation extremes everywhere, although a low signal to noise ratio may also play a role. However, even in regions where the Clausius-Clapeyron constraint is not closely followed, it still appears to be a better predictor for future changes in extreme precipitation than the change in mean precipitation (Pall et al., 2007). An observational study seems also to support this thermodynamical theory. Analysis of daily precipitation from the Special Sensor Microwave Imager (SSM/I) over the tropical oceans shows a direct link between rainfall extremes and temperature: heavy rainfall events increase during warm periods (El Niño) and decrease during cold periods (Allan and Soden, 2008). However, the observed amplification of rainfall extremes is larger than that predicted by climate models (Allan and Soden, 2008), due possibly to widely varying changes in upward velocities associated with precipitation extremes (O'Gorman and Schneider, 2008). Evidence from measurements in the Netherlands also suggest that hourly precipitation extremes may in some cases increase more strongly with temperature (twice as fast) than would be expected from the Clausius-Clapeyron relationship alone (Lenderink and Van Meijgaard, 2008), though this is still under debate (Haerter and Berg, 2009; Lenderink and van Meijgaard, 2009). A comparison between observed and multi-model simulated extreme precipitation using an optimal detection method suggests that the human-induced increase in greenhouse gases has contributed to the observed intensification of heavy precipitation events over large Northern Hemispheric land areas during the latter half of the 20th century (Min et al., 2011). One study has suggested that the influence of external forcing on daily extreme precipitation at regional scales would be detectable in another decade (Fowler and Wilby, 2010). However, this conclusion may be seasonally dependent. For example, by now there is about a 50% chance of detecting anthropogenic influence on UK extreme precipitation in winter, but the likelihood of the detection in other seasons is very small (Fowler and Wilby, 2010).

The present assessment based on evidence from new studies and those used in AR4 is that there is a *medium confidence* that anthropogenic influence has contributed to a trend towards increases in the frequency of heavy precipitation events over the 2nd half of the 20th century in many regions, especially in middle and higher latitudes of the Northern Hemisphere. This does not modify the AR4 assessment (see also Section 3.1.5). There is almost no literature on the attribution of changes in hail extremes, and thus no assessment can be provided for these at this point in time.

Regarding projected changes in extreme precipitation, the AR4 concluded that it was *very likely* that heavy precipitation events, i.e., the frequency of heavy precipitation or proportion of total precipitation from heavy precipitation, would increase over most areas of the globe in the 21st century (IPCC, 2007a). The tendency for an increase in heavy daily precipitation events was found in many regions including some regions in which the total precipitation was projected to decrease.

Post-AR4 analyses of climate model simulations partly confirm this assessment but also highlight fairly large uncertainties and model biases in projections of changes in heavy precipitation in some regions (Section 3.2.3 and Table 3.3). On the other hand, more GCM and RCM ensembles have now been analysed for some regions, leading to increased robustness of the projected changes (Table 3.3; see also e.g., Kharin et al., 2007; Kim et al., 2010; Hirschi et al., 2011). At the time of the AR4, Tebaldi et al. (2006) was the main global study available on projected changes in precipitation extremes (e.g., Figure 10.18 of Meehl et al., 2007b). Orlowsky and Seneviratne (2011) extended this analysis to a larger number of GCMs from the CMIP3 ensemble (see also Section 3.3.1). Figure 3.7 provides corresponding analyses of projected annual and seasonal changes of the wet-day intensity, the fraction of days with precipitation above the 95%-quantile of daily wet-day precipitation, and the fraction of days with precipitation above 10 mm/day. It should be noted that the 10 mm/day threshold cannot be considered extreme in several regions, but highlights differences in projections for absolute and relative thresholds (see also discussion in Section 3.1.2 and beginning of this section). All three indices were also considered in Tebaldi et al. (2006). Figure 3.7 indicates that regions with model agreement (at least 66%) with respect to changes in heavy precipitation are mostly found in the high latitudes and in the tropics, and in some mid-latitude regions of the Northern Hemisphere in the boreal winter. Regions with at least 90% model agreement are even more limited and confined to the high latitudes. Overall, model agreement in projected changes is found to be stronger in boreal winter (DJF) than summer (JJA) for most regions. Kharin et al. (2007) analyzed changes in annual maxima of 24-hour precipitation in the outputs of 14 CMIP3 simulations. Figure 3.8a displays the projected percentage change in annual maximum of 24-hour precipitation rate from the late-20th century 20-year return values, while Figure 3.8b displays the corresponding projected waiting times for late-20th-century 20-year return values of annual maximum 24-hour precipitation rates in the mid-21st century (left) and in late-21st century (right) under three different emission scenarios SRES B1, A1B and A2. Between the late 20th and the late 21st century, the projected responses of extreme precipitation to future emissions show increased precipitation rates in most regions, and decreases in waiting times in most regions in the high latitudes and the tropics and in some regions in the mid-latitudes (consistent with projected changes in heavy precipitation, see Figure 3.7 and Tebaldi et al., 2006), although there are increases in waiting times or only small changes projected in several regions (mostly in the southern half of South America, Central America, Central North America, Southern Asia, and Northern Australia). Except for these regions, the waiting period for an event of annual maximum 24-hour precipitation with a 20-year return period in the late-20th-century is projected to be about 5-15 years by the end of the 21st century. The greatest projected reductions in waiting time are in high latitudes and some tropical regions. The more extreme emissions scenarios (A1B and A2) lead to stronger projected decreases in waiting time.

[INSERT FIGURES 3.7 HERE

Figure 3.7: Projected annual and seasonal changes of three precipitation indices: Wet day intensity, fraction of days with precipitation above the 95%-quantile of daily wet day precipitation and fraction of days with pr > 10mm; CMIP3 simulations, 2080-2100 time frame minus 1980-2000 time frame (projections for A2 scenario, relative to late 20th century (20C3M) simulations), annual (top), DJF (middle) and JJA (bottom). Shading is only applied for areas where at least 66% of the models agree in the sign of the change; stippling is applied for regions where at least 90% of all models agree in the sign of the change [from Orlowsky and Seneviratne, 2011, after Tebaldi et al., 2006].]

[INSERT FIGURES 3.8 HERE

Figure 3.8:

(a, top) Projected changes from the late-20th-century 20-year return values of annual maximum 24-hour precipitation rates (%) in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the CMIP3, under three different SRES emission scenarios B1 (blue), A1B (green) and A2 (red) (adapted from Kharin et al., 2007).

The vertical extent of the whiskers shows the range of projected changes from all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Although the uncertainty range of projected change in extreme precipitation is large, the median model projection is that the extreme 24-hour precipitation rate will increase by about 5-10% by mid-21st century and by about 10-20% by late-21st century, depending on the region and the emissions scenario.
(b, bottom) Projected waiting times for late-twentieth-century 20-year return values of annual maximum 24-hour precipitation rates in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the CMIP3, under three different emission scenarios SRES B1 (blue), A1B (green) and A2 (red) (adapted from Kharin et al., 2007). The vertical extent of the whiskers in both directions describes the range of projected changes by all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Although the uncertainty range of projected change in extreme precipitation is large, almost all models suggest that the waiting time for a late 20th century 20-year extreme 24-hour precipitation event will be reduced to substantially less than 20 years by mid-21st and much more by late-21st century, indicating an increase in frequency of the extreme precipitation at continental and sub-continental scales under all three forcing scenarios. Two global domains for which projections are shown are: the entire globe including the oceans, and the global land areas.]

Future precipitation projected by the CMIP3 models has also been analyzed in a number of studies for various regions using different combinations of the models (Table 3.3 and next paragraphs). In general these studies confirm the findings of global-scale studies by Tebaldi et al. (2006), Kharin et al. (2007) and Orlowsky and Seneviratne (2011).

By analyzing simulations with a single GCM, Khon et al. (2007) reported a projected general increase in extreme precipitation for the different regions in northern Eurasia especially for winter. Su et al. (2009) found that for the Yangtze River Basin region in 2001–2050, the 50-year heavy precipitation and drought events become more frequent, with return periods falling to below 25 years (relative to 1951-2000 behavior). For the Indian region, the Hadley Centre coupled model HadCM3 projects increases in the magnitude of the heaviest rainfall with CO_2 doubling (Turner and Slingo, 2009). Simulations by 12 GCMs projected an increase in heavy precipitation intensity and mean precipitation rates and less severe droughts in east Africa, more severe precipitation deficits in the southwest of southern Africa, and enhanced precipitation farther north in Zambia, Malawi, and northern Mozambique (Shongwe et al., 2009, 2011). Rocha et al. (2008) evaluated differences in the precipitation regime over southeastern Africa simulated by two GCMs under present (1961–1990) and future (2071–2100) conditions as a result of greenhouse gases anthropogenic forcing. They found that the intensity of all episode categories of precipitation events is projected to increase practically over the whole region, whereas the number of episodes is projected to decrease in most of the region and for most episode categories. Extreme precipitation is projected to increase over Australia in 2080–2099 relative to 1980–1999 in an analysis of the CMIP3 ensemble (Alexander and Arblaster, 2009). In addition, several high-spatial resolution studies are available in different regions, as highlighted in the following paragraph.

High-spatial resolution is important for studies of extreme precipitation (e.g., Kim et al., 2010). Post-AR4 studies have employed three approaches to obtain high-spatial resolution to project precipitation extremes: high-resolution GCMs, dynamical downscaling using RCMs, and statistical downscaling. Kamiguchi et al. (2006) is an example of studies that employed the first approach. With the Meteorological Research Institute and Japan Meteorological Agency (MRI-JMA) 20-km horizontal grid AGCM that was run in time slice mode, heavy precipitation was projected to increase substantially in south Asia, the Amazon, and west Africa, with increased dry spell persistence in South Africa, southern Australia, and the Amazon at the end of the 21st century. In the Asian monsoon region, heavy precipitation was projected to increase, notably in Bangladesh and in the Yangtze River basin due to the intensified convergence of water vapor flux in summer. Using statistical downscaling, Wang and Zhang (2008) investigated possible changes in North American extreme precipitation probability during winter from 1949–1999 to 2050–2099. Downscaled results suggested a strong increase in extreme precipitation over the south and central U.S. but decreases over the Canadian prairies. Projected European precipitation extremes in high-resolution studies tend to increase in northern Europe (Frei et al., 2006; Beniston et al., 2007; Schmidli et al., 2007), especially during winter (Haugen and Iversen, 2008; May, 2008), as also highlighted in Table 3.3. Fowler and Ekström (2009) project increases in both short-duration (1-day) and longer-duration (10-day) precipitation extremes across the UK during winter, spring and autumn. In summer, model projections for the UK span the zero change line, although there is low confidence due to poor model performance in this season. Using daily statistics from various models, Boberg et al. (2009a, b) projected a clear increase in the contribution to total precipitation from more intense events together with a decrease in the number of days with light precipitation. This pattern of change was found to be robust for all European sub-regions. In double-nested model simulations with a horizontal grid spacing of 10 km, Tomassini and Jacob (2009) projected positive trends in extreme quantiles of heavy precipitation over Germany, although they are relatively small compared with the uncertainties except for the higher emissions A2 scenario. For the Upper Mississippi River Basin region during October–March, the intensity of extreme precipitation is projected to increase (Gutowski et al., 2008b). Simulations with a single RCM project an increase in the intensity of extreme precipitation events over most of southeastern South America and western Amazonia in 2071–2100, whereas in northeast Brazil and eastern Amazonia smaller or no changes are projected (Marengo et al., 2009a). Outputs from another RCM indicate an increase in the magnitude of future extreme rainfall events in the Westernport region of Australia, consistent with results based on the CMIP3 ensemble (Alexander and Arblaster, 2009), and the size of this increase is greater in 2070 than in 2030 (Abbs and Rafter, 2008). When both future land use changes and increasing greenhouse-gas concentrations are considered in the simulations, tropical and northern Africa are projected to experience less extreme rainfall events by 2025 during most seasons except for autumn (Paeth and Thamm, 2007). Simulations with high resolution RCMs projected that frequency of extreme precipitation increases in the warm climate for warm or rainy season in Japan (Nakamura et al., 2008; Wakazuki et al., 2008; Kitoh et al., 2009). An increase in 90th-percentile values of daily precipitation on the Pacific side of the Japanese Islands during July in the future climate was projected with a 5-km mesh cloud-system resolving non-hydrostatic RCM (Kanada et al., 2010b).

In summary, projected changes from both global and regional studies indicate that it is *likely* that the frequency of heavy precipitation (or proportion of total rainfall from heavy falls) will increase in the 21st century over many areas in the globe, especially in the high latitudes and tropical regions, and northern mid-latitudes in winter (Table 3.3 and Figures 3.7 and 3.8). This represents a weaker (*likely* versus *very likely*) but more robust assessment than that of the AR4 (Meehl et al., 2007b), as it is based on a larger number of studies based on more numerous lines of evidence.

Post-AR4 studies indicate that the projection of precipitation extremes is associated with large uncertainties, contributed by the uncertainties related to GCMs, RCMs and statistical downscaling methods, and by the impacts of natural variability of the climate. Kysely and Beranova (2009) examined scenarios of change in extreme precipitation events in 24 future climate runs of 10 RCMs driven by two GCMs, focusing on a specific area of central Europe with complex orography. They demonstrated that the inter- and intra-model variability and related uncertainties in the pattern and magnitude of the change are large, although they also show that the projected trends tend to agree with those recently observed in the area, which may strengthen their credibility. May (2008) reported an unrealistically large projected precipitation change over the Baltic Sea in summer in the HIRHAM RCM, apparently related to an unrealistic projection of Baltic Sea warming in the driving GCM. Frei et al. (2006) found large model differences in summer when RCM formulation contributes significantly to scenario uncertainty. In exploring the ability of two statistical downscaling models in reproducing the direction of the projected changes in indices of precipitation extremes Hundecha and Bardossy (2008) concluded that statistical downscaling seems to be more reliable during seasons when local climate is determined by large-scale circulation than by local convective processes. Themeßl et al. (2011) merged linear and nonlinear empirical-statistical downscaling techniques with bias correction methods, and demonstrated their ability to drastically reduce RCM error characteristics. The extent to which the natural variability of the climate affects our ability to project the anthropogenically forced component of changes in daily precipitation extremes was investigated by Kendon et al. (2008). They show that annual to multidecadal natural variability across Europe may contribute to substantial uncertainty. Also, Kiktev et al. (2009) performed an objective comparison of climatologies and historical trends of temperature and precipitation extremes using observations and 20th century climate simulations. They did not detect significant similarity between simulated and actual patterns for the indices of precipitation extremes in most cases. Moreover, Allan and Soden (2008) used satellite observations and model simulations to examine the response of tropical precipitation events to naturally driven changes in surface temperature and atmospheric moisture content. The observed amplification of rainfall extremes was larger than that predicted by models. The underestimate of rainfall extremes by the models may be related to the coarse spatial resolution used in the model simulations and suggests that projections of future changes in rainfall extremes in response to anthropogenic global warming may be underestimated.

Confidence is still low for hail projections particularly due to a lack of hail-specific modelling studies, and a lack of agreement among the few available studies. There is little information in the AR4 regarding projected changes in hail events, and there has been little new literature since the AR4. Leslie et al. (2008) used coupled climate model simulations under the SRES A1B scenario to estimate future changes in hailstorms in the Sydney Basin, Australia. Their future climate simulations show an increase in the frequency and intensity of hailstorms out to 2050, and they suggest that the increase will emerge from the natural background variability within just a few decades. This result offers a different conclusion from the modelling study of Niall and Walsh (2005), which simulated Convective Available Potential Energy (CAPE) for southeastern Australia in an environment containing double the pre-industrial concentrations of equivalent CO_2. They found a statistically significant projected decrease in CAPE values and concluded that “it is possible that there will be a decrease in the frequency of hail in southeastern Australia if current rates of CO_2 emission are sustained”, assuming the strong relationship between hail incidence and the CAPE for 1980-2001 remains unchanged under enhanced greenhouse conditions.

In summary, it is *likely* that there has been statistically significant increases in the number of heavy precipitation events (e.g., 95th percentile) in more regions than there has been statistically significant decreases, but there are strong regional and subregional variations in the trends. There is *medium confidence* that changes in extreme precipitation at global scale may have been anthropogenically related. It is *likely* that the frequency of heavy

precipitation (or proportion of total rainfall from heavy falls) will increase in the 21st century over many areas, in particular in the high latitudes and tropical regions, and northern mid-latitudes in winter (Table 3.3 and Figures 3.7 and 3.8). Some studies also suggest increases in heavy precipitation in some regions with projected decreases of total precipitation, such as Central Europe (*medium confidence*). A one-in-20 year annual maximum 24-hour precipitation rate could *likely* become a one in 5- to 15-year event by the end of 21st century in many regions, but some regions display decreases or statistically non-significant changes in heavy precipitation based on current climate model projections.

3.3.3. Wind

Extreme wind speeds pose a threat to human safety, maritime and aviation activities and the integrity of infrastructure. As well as extreme wind speeds, other attributes of wind can cause extreme impacts. Trends in average wind speed can influence evaporation and in turn water availability and droughts (e.g., McVicar et al., 2008). Sustained mid-latitude winds can elevate coastal sea levels (e.g., McInnes et al., 2009b) while longer term changes in prevailing wind direction can cause changes in wave climate and coastline stability (Pirazzoli and Tomasin, 2003; see also Section 3.5.4 and 3.5.5). Aeolian processes exert significant influence on the formation and evolution of arid and semi-arid environments, being strongly linked to soil and vegetation change (Okin et al., 2006). A rapid shift in wind direction may reposition the leading edge of a forest fire (see Section 4.2.2.2, Mills, 2005) while the fire itself may generate a local circulation response such as tornadogenesis (e.g., Cunningham and Reeder, 2009). Unlike other weather and climate elements such as temperature and rainfall, extreme winds are often considered in the context of the extreme phenomena with which they are associated such as tropical and extratropical cyclones (see also Sections 3.4.4 and 3.4.5), thunderstorm downbursts and tornadoes. Changes in wind extremes may arise from changes in the intensity or location of their associated phenomena or from other changes to the climate system (e.g., a change in local convective activity). Although wind is often not used to define the extreme event itself (Peterson et al., 2008c), wind speed thresholds may be used to characterize the severity of the phenomenon (e.g., the Saffir-Simpson scale for tropical cyclones).

Changes in wind climate over paleo-climatic time scales were not addressed specifically in the AR4 but may be inferred from circulation changes determined from reconstructions using proxy data. Broad circulation changes have occurred across the globe from the mid-Holocene (~ 6000 years ago) to the beginning of the industrial revolution (Wanner et al., 2008). Over this period, there was a change toward a lower Northern Atlantic Oscillation (NAO) index, implying weaker westerly winds over the north Atlantic, and the ITCZ moved southward leading to weaker monsoons across Asia. The Walker Circulation strengthened, El Niño activity was higher and Southern Ocean westerlies moved northward and strengthened affecting southern Australia, New Zealand and southern South America. While the changes in the Northern Hemisphere corresponded to changes in orbital forcing, those in the Southern Hemisphere were more complex, possibly reflecting the additional role on circulation of heat transport in the ocean. Solar variability and volcanic eruptions may also have contributed to decadal to multi-centennial fluctuations over this time period (Wanner et al., 2008).

The AR4 did not specifically address changes in extreme wind although it did report on wind changes in the context of other phenomena such as tropical and extratropical cyclones and oceanic waves and concluded that mid-latitude westerlies had increased in strength in both hemispheres (Trenberth et al., 2007). Long-term high-quality wind measurements from terrestrial anemometers are sparse in many parts of the globe due to the influence of changes in instrumentation, station location, and surrounding land use (e.g., Cherry, 1988; Pryor et al., 2007; Jakob, 2011), and this has hampered the direct investigation of wind climatology changes. Nevertheless a number of recent studies have analysed mean and extreme wind speed trends from wind observations in different parts of the world. Wan et al. (2010) used a long-term (1953-2006) data series of 1-minute-mean near-surface hourly data standardised to 10 m and found decreasing trends in monthly averaged winds over western and most parts of southern Canada (except the Maritimes) in all seasons, with significant increases in the central Canadian Arctic in all seasons and in the Maritimes in spring and autumn. Over the Gulf of St Lawrence, Hundecha et al. (2008) found declining trends in annual maximum winds in the north of the gulf and increasing trends to the south in North American Regional Reanalysis (NARR) data and thirteen anemometer sites over the period 1979–2004 although the changes were mostly not statistically significant. Pryor et al. (2007) reported mostly declining trends in wind over much of the USA in 50th and 90th percentile wind speeds calculated from time series of twice daily winds at 157 sites over 1973 to 2005. Lynch et al. (2004) found statistically significant increasing trends in average winds but reported that no trends were found for the highest winds (highest daily average wind speed reported per season) in Alaska from 1955-2001. However, they note that instrument and measurement changes may have influenced their conclusions. Pirazolli and Tomasin (2003) reported a generally declining trend in both annual mean and annual maximum winds from 1951 to the mid-1970s and an increasing trend since then, based on central Mediterranean records. Over the Netherlands, Smits et al. (2005) found declining trends in winds occurring on average 10 and 2 times per year in 10-m anemometer data over 1962-2002 but increasing trends in NCEP and ERA40 reanalysis. Over China, negative trends in 10-m winds were found by Guo et al. (2011) based on winds from 652 sites over 1969-2005, and by Jiang et al. (2010a) in daily maximum wind speeds based on 535 sites over 1956–2004 and by Zhang et al. (2007b) in 2 m data at 75 sites from 1966-2003 over the Tibetan plateau, consistent with an earlier study by Xu et al. (2006) who also found declining trends in both station data and NCEP reanalyses.

Strong winds, defined by Jiang et al. (2010a) as 10 minute average exceedences or daily average exceedences of 17 m/s, and by Guo et al. (2011) as high percentiles, also declined over the period mainly during the spring and at similar rates at urban and rural sites. McVicar et al. (2008) using 2 m wind data over the 1975-2006 period reported declines in mean wind speed over 88% of Australia (significant over 57% of the country) and positive, though not necessarily significant, trends over about 12% of the mainland interior and southern and eastern coastal regions including Tasmania, whereas mostly increasing trends were found in both NCEP and ERA40 10-m winds. However, Troccoli et al. (2011) found these trends were highly sensitive to the measurement elevation, confirming the declining trend at 2 m but finding mostly increasing trends at 10 m in broad agreement with the reanalysis data over 1975-2006. In Antarctica, Turner et al. (2005) reported increasing trends in mean wind speeds over the second half of the 20th century. Consistent with many of the northern European studies, Vautard et al. (2010) found mostly declining trends across most of the continental northern mid-latitudes. Another recent study by McVicar et al. (2010) suggests, based on observations (1960-2006) from mountainous regions in Switzerland and China, that mean near-surface wind speeds are possibly declining more rapidly at higher elevations than lower elevations in these areas. Using a new dataset of ship-based anemometer wind over the period 1950-2008, positive wind speed trends were found over the Indian Ocean, extending through Indonesia to the northern Pacific while negative trends were found over the Central Pacific (Tokinaga and Xie, 2011). However trends in winds over 1987-2006 differed in sign from satellite derived winds around Australia and parts of southeast Asia.

Proxies for wind that use pressure tendencies and geostropic winds calculated from triangles of pressure observations from which storminess can be inferred have also been employed in a number of studies over Europe and the Atlantic (see 3.4.5). These studies suggest that there was a tendency for increased storminess around 1900 and in the 1990s, while the 1960s and 1970s were periods of low storm activity; but there are no long-term trends consistent between different available studies. More recent studies confirm these findings and indicate that storminess in this region exhibits strong inter-decadal variability (Alexandersson et al., 2000; Allan et al., 2009; Wang et al., 2009c). The latter half of the 20th century was punctuated by a peak in storminess around 1990 which according to Wang et al. (2009c) is unprecedented since 1874. However, no long-term trends were detected in storminess over this time period (Barring and von Storch, 2004; Barring and Fortuniak, 2009) or the period for which reanalysis data exist (Raible, 2007; Della-Marta et al., 2009). No statistically significant trends have been detected in the global annual number of tropical cyclones to date although a trend has been detected in the intensity of the strongest storms since 1980 (see 3.4.3). Regarding other phenomena associated with extreme winds, studies on thunderstorms, tornadoes and mesoscale convective complexes are too few in number to be used to infer extreme wind speed change.

The AR4 did not address the causes of changes in extreme winds but reported that anthropogenic forcing is *likely* to have contributed to changes in wind patterns, affecting extratropical storm tracks in both hemispheres although it was noted that the observed changes in the Northern Hemisphere circulation are larger than those simulated in response to 20th-century forcing change. The relationship between mean and severe winds and natural modes of variability has been investigated in several post-AR4 studies. On the British Columbian coast, Abeysirigunawardena et al. (2009) found that higher extreme winds tend to occur during the negative (i.e., cold) ENSO phase. The generally increasing trend of mean wind speeds over recent decades in Antarctica is consistent with the change in the nature of the Southern Annular Mode towards its high index state (Turner et al., 2005). Donat et al. (2010b) concluded that 80% of storm days in Central Europe are connected with westerly flows which occur primarily during the positive phase of the NAO. Declining trends in wind over China have mainly been linked to circulation changes due to a weaker land-sea thermal contrast (Xu et al., 2006; Jiang et al., 2010a; Guo et al., 2011). Vautard et al. (2010) attribute the slow down in surface winds over most of the continental northern mid-latitudes to changes in atmospheric circulation (10-50%) and an increase in surface roughness due to biomass increases (25-60%) which are supported by regional climate model simulations. Wang et al. (2009d), formally detected a link between external forcing and positive trends in the high northern latitudes and negative trends in the northern mid-latitudes using a proxy for wind (geostrophic wind energy) in the boreal winter. Trends in the central Mediterranean were found to be positively correlated with temperature but not with the NAO index (Pirazzoli and Tomasin, 2003).

Projections of wind speed changes in general and wind extremes in particular were not specifically addressed in the AR4 although references are made to wind speed in relation to other variables and phenomena such as mid-latitude storm tracks, tropical cyclones and ocean waves (Christensen et al., 2007; Meehl et al., 2007b). The AR4 (2007a) reported that it was likely that future tropical cyclones (typhoons and hurricanes) would become more intense, with larger peak wind speeds associated with ongoing increases of tropical SSTs. It also reported that there was higher confidence in the projected poleward shift of the storm tracks and associated changes in wind patterns. Since the AR4 there have been several studies which have focussed on future changes to extreme winds. Gastineau and Soden (2009) found agreement between models of a decreased frequency in the tropics and increased frequency in the extratropics of the strongest wind events based on changes in percentiles of 850 hPa wind speed, that are expected to be representative of winds at the surface, using a 17-model ensemble. This is consistent with changes found in 10 m winds by McInnes et al. (2011) in 99th percentile wind speed using daily wind speeds from 19 models from the CMIP3 ensemble. Results from that study, showing changes in mean and 99th percentile winds for 2081-2100 relative to 1981-2000, are shown in Figure 3.9 to illustrate the degree of spatial agreement between the CMIP3 ensemble regarding wind speed change. In

DJF, there is general model agreement on mean wind speed increase over Europe, the northern Pacific, northeastern Canada, north Africa and the Southern Ocean south of 45ºS. Agreement on mean wind speed decrease occurs over the Mediterranean, northeastern to southwestern Indian Ocean and the Southern Hemisphere between 35 and 45ºS. In JJA, model agreement on mean wind speed increase occurs over Europe and the Mediterranean, the tropical Pacific, northern Australia and Indian Ocean between about 10 and 30ºS and the southern Ocean south of 40ºS. Mean wind speed decrease occurs over parts of the Pacific and Indian Oceans from the Equator northwards and in the southern hemisphere at 30-40ºS. Extreme wind speeds (bottom panels of Figure 3.9) show consistency between models over a large portion of the globe in the direction of change, which are in many areas consistent with the direction of change of the mean winds. Increases in extremes occur in the high latitudes of both hemispheres and decreases across the lower latitudes as reported by Gastineau and Soden (2009), but regional differences are apparent. For example, in DJF, consistent increases in extremes are seen across much of northern Europe, north Africa and parts of Asia and eastern North America. In JJA, agreement on increases in extremes of up to 5% are seen in eastern South America, northern Australia, the south Pacific between 10 and 20ºS, parts of Africa and much of the Southern Ocean, while agreement on decrease is seen across large parts of the Atlantic and Indian Oceans, the northeast Pacific and northern North America. In some areas such as eastern Asia in DJF, models agree on declining mean winds together with an increase in extreme winds whereas over the south Pacific, southern Australia and Indian Ocean at around 30ºS models agree on an increasing trend in mean winds together with declining trends in extreme winds. While these maps indicate where models produce consistent changes in mean and extreme winds, it should be noted that high agreement across GCMs on a particular sign of extreme wind change may not necessarily indicate a more reliable result because models at their current resolution are unable to resolve small scale phenomena such as tropical cyclones, tornadoes and mesoscale convective complexes that are associated with particularly severe winds and this lowers the confidence in the extreme wind changes particularly in the regions most influenced by these phenomena. For instance, Diffenbaugh et al. (2008) noted that increased atmospheric greenhouse gas concentrations may cause some of the atmospheric conditions conducive to tornadoes such as atmospheric instability to increase due to increasing temperature and humidity, while others such as vertical shear to decrease due to reduced pole-to-equator temperature gradient. They concluded that this limited confidence in the sign of any possible change in tornado activity.

[INSERT FIGURE 3.9 HERE

Figure 3.9: The average of the projected multi-model 10 m mean wind speeds (top) and 99th percentile daily wind speeds (bottom) for the period 2080 to 2099 relative to 1980 to 1999 (% change) for December to February (left) and June to August (right) plotted only where more than 66% of the models agree on the sign of the change. Fine black stippling indicates where more than 90% of the models agree on the sign of the change and bold grey stippling (in white or light coloured areas) indicates where 66% of models agree on a small change between ±2 %. From McInnes et al., (2011).]

In addition to the global studies, several regional studies of projected winds have also been undertaken and while these provide detail over specific regional areas there is still considerable variation in results related to the different models and methods used. Debernard and Roed (2008) projected statistically significant increases in 99th percentile winds over 2071-2100 relative to 1961-1990 across much of northern Europe, the British Isles and the ocean to the west and decreases to the south of Iceland in several models under various SRES emission scenarios (A2, B2, A1B). Rockel and Woth (2007) projected an increase in winter and a decrease in autumn of 99th percentile daily mean wind speed from eight RCMs over 2071-2100 relative to 1961-1990, over European areas influenced by North Atlantic extra-tropical cyclones. Donat et al. (2010a; 2011) reported an increase of between 19 and 33% of wind storm days in winter over large parts of Central Europe, associated with an increase in intensity of cyclones and a 5% increase in associated winds for the end of the 21st Century relative to the late 20th Century in an ensemble of 9 GCMs although the changes were not statistically significant in all models. Haugen and Iversen (2008) dynamically downscaled four GCMs using an RCM and noted that daily maximum wind speeds in the future climate become more frequent over large parts of northern Europe. Beniston et al. (2007) used 9 RCMs to downscale two GCMs and found that extreme wind speeds increased and become more northwesterly between 45° and 55°N, except over and south of the Alps, for 2071-2100 relative to 1961-1990, although the magnitude of the increase depends on the specific RCM used. These changes were attributed to reductions in mean sea-level pressure and the generation of more North Sea storms. In contrast, Sterl et al. (2009) using a 17 member ensemble of a single GCM found an increase in strong winds from the southwest rather than the north. Leckebusch et al (2008) used storm indices applied to GCM simulations and found that high wind speeds in the future climate increased due to both an increase in the storm wind speed magnitude and the length of the storm tracks. Rauthe et al. (2010) dynamically downscaled ECHAM5 at better than 20 km resolution over the period 2021-2050 relative to 1971-2000 for A1B, A2 and B1 using two RCMs and reported a change in the 10-year wind speed of +6 to -1.5 % over northern Germany and a slight decrease in the majority of simulations over central and southern Germany. Pinto et al. (2007a) used an ensemble of simulations from a GCM forced with A1B and A2 scenarios to estimate changes to insured loss potential (a function of excesses of 98th percentile wind and population density) at the end of the 21st century and found increases everywhere with largest changes in Germany and France, and smallest changes for Portugal and Spain. Over China, Jiang et al. (2010b) projected decreases in winter and annual mean wind speed for 2081-2100 compared to 1971-1990 based on two RCMs that downscale two different GCMs. Sailor et al. (2008) statistically downscaled winds from four GCMs, to develop projections of winds over five airports in the northwest U.S. and the results for 2050 suggested that summertime wind speeds might decrease by 5–10%, while changes to wintertime wind speeds were less certain.

In summary, post-AR4 studies provide considerable new evidence of changes in wind speed across the globe. A declining trend has been detected in mean wind speed across much of the Northern Hemisphere continents. In one study this has been attributed in part to circulation changes and in part to increasing surface roughness. However, due to the various shortcomings associated with anemometer data and the inconsistency in anemometer and reanalysis trends in some regions, we have *low confidence* in the causes of the trends at this stage. We have also *low confidence* in how the observed trends in mean wind speed relate to trends in strong winds. The relatively few studies of projected extreme winds, combined with the different models, regions and methods used to develop projections of this quantity, means that we have *low confidence* in projections of changes in strong winds.

3.4. Observed and Projected Changes in Phenomena Related to Weather and Climate Extremes

3.4.1. Monsoons

Changes in monsoon-related extreme precipitation and winds due to climate change are not well understood, but a variety of extremes such as floods, drought or even heatwaves may occur more or less frequently than present in the monsoon regions as a consequence of climate change. Generally, precipitation is the most important variable for inhabitants of monsoon regions, but it is also a variable associated with larger uncertainties in climate simulations (Wang et al., 2005; Kang and Shukla, 2006). Changes in the monsoons can be characterized more broadly than via precipitation only, thus monsoon changes might be better depicted by large-scale dynamics, circulation or moisture convergence. However, few studies have focused on observed changes in the large-scale and regional monsoon circulations. Hence, in this section we focus on monsoon-induced changes in rainfall, but when literature is available we also provide assessments on associated circulation changes. The focus in this section is also on total and seasonal rainfall, with most discussions of intense rainfall covered in 3.3.2.

Modeling experiments to assess paleo-monsoons suggest that in the past, during the Holocene due to orbital forcing on a millennial timescale, there was a progressive southward shift of the Northern Hemisphere (NH) summer position of the Intertropical Convergence Zone (ITCZ). This was accompanied by a pronounced weakening of the monsoon systems in Africa and Asia and increasing dryness on both continents, while in South America the monsoon was weaker, as suggested both by models and paleo indicators (Wanner et al., 2008).

The delineation of the global monsoon has been mostly performed using rainfall data or outgoing longwave radiation (OLR) fields (Kim et al., 2008). Zhou et al. (2008b; 2008a) and Wang and Ding (2006) report that the combination of monsoon area and rainfall intensity change has led to an overall weakening trend of global land monsoon rainfall accumulation during the last 54 years. Lau and Wu (2007) identified two opposite time evolutions in the occurrence of rainfall events in the tropics, in overall agreement with the Climate Research Unit's gauge-only rainfall data over land: a negative trend in moderate rain events and a positive trend in heavy and light rain events. Positive trends in intense rain were located in deep convective cores of the ITCZ, South Pacific Convergence Zone, Indian Ocean and monsoon regions.

In the Indo-Pacific region, covering the southeast Asian and north Australian monsoon, Caesar et al. (2011) found low spatial coherence in trends in precipitation extremes across the region between 1971 and 2003. In the few cases where statistically significant trends in precipitation extremes were identified, there was generally a trend towards wetter conditions, in common with the global results of Alexander et al. (2006). Liu et al. (2011) reported a decline in recorded precipitation events in China 1960–2000, which was mainly accounted for by a decrease of light precipitation events, with intensities of 0.1–0.3 mm/day. Some of the extreme precipitation appeared to be positively correlated with a La Niña-like SST pattern, but without suggesting the presence of a trend. With regard to wind changes, Guo et al. (2011) analyzed near-surface wind speed change in China and its monsoon regions from 1969 to 2005 and showed a statistically significant weakening in annual and seasonal mean wind.

For the Indian monsoon, Rajeevan et al. (2008) showed that extreme rain events have an increasing trend between 1901 and 2005, but the trend is much stronger after 1950. Sen Roy (2009) investigated changes in extreme hourly rainfall in India, and found widespread increases in heavy precipitation events across India, mostly in the high-elevation regions of the northwestern Himalaya as well as along the foothills of the Himalaya extending south into the Indo-Ganges basin, and particularly during the summer monsoon season during 1980-2002.

In the African monsoon region, Fontaine et al. (2011) investigated recent observed trends using high-resolution gridded precipitation from the CRU (period 1979–2002), OLR and the NCEP reanalyses. Their results revealed a rainfall increase in north Africa since the mid-90s. Over the longer term, however, Zhou et al. (2008b; 2008a) and Wang and Ding (2006) reported an overall weakening trend of global land monsoon rainfall accumulation during the last 54 years, which was mainly caused by the North African monsoon and South Asian monsoon.

For the North American monsoon region, Cavazos et al. (2008) reported increases in the intensity of precipitation in the mountain sites of northwestern Mexico over the 1961-1998 period, apparently related to an increased contribution from heavy precipitation derived from tropical cyclones. Arriaga-Ramirez and Cavazos (2010) found that total and extreme rainfall in the monsoon region of western Mexico and the U.S. southwest presented a statistically significant increase during 1961–1998, mainly in winter. Groisman and Knight (2008) found that consecutive dry days (see Box 3.2 for definition) with periods longer than one month have significantly increased in the U.S. southwest. Increases in heavy precipitation during 1960-2000 in the South American monsoon have been documented by Marengo et al. (2009a; 2009b), and Rusticucci et al. (2010). Studies using circulation fields such as 850 hPa winds or moisture flux have been performed for the South American monsoon system for assessments of the onset and end of the monsoon, and indicate that the onset exhibits a marked interannual variability linked to variations in SST anomalies in the Eastern Pacific and tropical Atlantic (Gan et al., 2006; da Silva and de Carvalho, 2007; Raia and Cavalcanti, 2008; Nieto-Ferreira and Rickenbach, 2011).

Attributing the causes of changes in monsoons is difficult because there are substantial inter-model differences in representing Asian monsoon processes (Christensen et al., 2007). Most models simulate the general migration of seasonal tropical rain, although the observed maximum rainfall during the monsoon season along the west coast of India, the North Bay of Bengal and adjoining northeast India is poorly simulated by many models. Bollasina and Nigam (2009) show the presence of large systematic biases in coupled simulations of boreal summer precipitation, evaporation, and SST in the Indian Ocean, often exceeding 50% of the climatological values. Many of the biases are pervasive, being common to most simulations.

The observed negative trend in global land monsoon rainfall is better reproduced by atmospheric models forced by observed historical sea surface temperature (SST), than by coupled models without explicit forcing by observed ocean temperatures (Kim et al., 2008). This trend is strongly linked to the warming trend over the central eastern Pacific and the western tropical Indian Ocean (Zhou et al., 2008b). For the west African monsoon, Joly and Voldoire (2010) explore the role of Gulf of Guinea SSTs in its interannual variability. In most of the studied CMIP3 simulations, the inter-annual variability of SST is very weak in the Gulf of Guinea, especially along the Guinean Coast. As a consequence, the influence on the monsoon rainfall over the African continent is poorly reproduced. It is suggested that this may be due to the counteracting effects of the Pacific and Atlantic basins over the last decades. The decreasing long-term trend in north African summer monsoon rainfall may be due to the atmosphere response to observed SST variations (Hoerling et al., 2006; Zhou et al., 2008b; Scaife et al., 2009). A similar trend in global monsoon precipitation in land regions is reproduced in CMIP3 models' 20th century simulations when they include anthropogenic forcing, and for some simulations natural forcing (including volcanic forcing) as well, though the trend is much weaker in general, with the exception of one model (HadCM3) capable of producing a trend of similar magnitude (Li et al., 2008). The decrease in east Asian monsoon rainfall also seems to be related to tropical SST changes (Li et al., 2008), and the less spatially coherent positive trends in precipitation extremes in the southeast Asian and north Australian monsoons appear to be positively correlated with a La Niña-like SST pattern (Caesar et al., 2011).

A variety of factors, natural and anthropogenic, have been suggested as possible causes of variations in monsoons. Changes in regional monsoons are strongly influenced by the changes in the states of dominant patterns of climate variability such as the El Niño – Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO), the Northern Annular Mode (NAM), the Atlantic Multi-decadal Oscillation (AMO), and the Southern Annular Mode (SAM) (see also Sections 3.4.2 and 3.4.3). Additionally, model-based evidence has suggested that land surface processes and land use changes could in some instances significantly impact regional monsoon. Tropical land cover change in Africa and southeast Asia appears to have weaker local climatic impacts than in Amazonia (Voldoire and Royer, 2004; Mabuchi et al., 2005a, b). Grimm et al. (2007) and Collini et al. (2008) explored possible feedbacks between soil moisture and precipitation during the early stages of the monsoon in South America, when the surface is not sufficiently wet, and soil moisture anomalies may thus also modulate the development of precipitation. However, the influence of historical land use on monsoon is difficult to quantify, due both to the poor documentation of land use and difficulties in simulating monsoon at fine scales. The impact of aerosols (black carbon and sulfate) on monsoon regions has been discussed by Meehl et al. (2008), Lau et al. (2006) and Silva Dias et al. (2002). These studies suggest that there are still large uncertainties and a strong model dependency in the representation of the relevant land surface processes and the role of aerosol direct forcing, and resulting interactions (e.g., in the case of land use forcing; Pitman et al., 2009).

Regarding projections of change in the monsoons, the AR4 concluded (Christensen et al., 2007) that there "is a tendency for monsoonal circulations to result in increased precipitation due to enhanced moisture convergence, despite a tendency towards weakening of the monsoonal flows themselves. However, many aspects of tropical climatic responses remain uncertain." As global warming is projected to lead to faster warming over land than over the oceans (e.g., IPCC, 2007a; Sutton et al., 2007), the continental-scale land-sea thermal contrast, a major factor affecting monsoon circulations, may become stronger in summer. Based on this hypothesis, a simple scenario is that the summer monsoon will be stronger and the winter monsoon will be weaker in the future than the present. However, model results derived from the analyses of 15 CMIP 3 global models are not as straightforward as this simple consideration (Tanaka et al., 2005), as they show a weakening of these tropical circulations by the late 21st century compared to the late 20th century. In turn, such changes in circulation may lead to changes in precipitation associated with monsoons. For instance, the monsoonal precipitation in Mexico and Central America is projected to decrease in association with increasing precipitation over the eastern equatorial Pacific through changes in the Walker Circulation and local Hadley Circulation (e.g., Lu et al., 2007). Complicating this picture further, however, is the fact that observations and models suggest that changes in monsoons are related at least in part to changes in observed SSTs, as noted above.

At regional scales, there is little consensus in GCM projections regarding the sign of future change in the monsoons characteristics, such as circulation and rainfall. For instance, while some models project an intense drying of the Sahel under a global warming scenario, others project an intensification of the rains, and some project more frequent extreme events (Cook and Vizy, 2006). Increases in precipitation are projected in the Asian monsoon (along with an increase in interannual season-averaged precipitation variability), and in the southern part of the west African monsoon, but with some decreases in the Sahel in northern summer. In the Australian monsoon in southern summer, an analysis by Moise and Colman (2009) from the entire ensemble mean model of CMIP3 simulations suggested no changes in Australian tropical rainfall during the summer and only slightly enhanced inter-annual variability.

A study of 19 CMIP3 models indicates a significant increase in mean south Asian summer monsoon precipitation of 8% and a possible extension of the monsoon period, together with intensification of extreme excess and deficient monsoons (Kripalani et al., 2007). A more recent study (Ashfaq et al., 2009) from the downscaling of the NCAR CCSM3 global model using the RegCM3 regional model suggests a weakening of the large-scale monsoon flow and suppression of the dominant intraseasonal oscillatory modes with overall weakening of the south Asian summer monsoon by the end of the 21st century resulting in a decrease in summer precipitation in key areas of south Asia.

Kitoh and Uchiyama (2006) used 15 models under the A1B scenario to analyze the changes in intensity and duration of precipitation in the Baiu-Changma-Meiyu rain band at the end of the 21st century. They found a delay in early summer rain withdrawal over the region extending from Taiwan, Ryukyu Islands to the south of Japan, contrasted with an earlier withdrawal over the Yangtze Basin. They attributed this feature to El Niño-like mean state changes over the monsoon trough and subtropical anticyclone over the western Pacific region. A southwestward extension of the subtropical anticyclone over the northwestern Pacific Ocean associated with El Niño-like mean state changes and a dry air intrusion at the mid-troposphere from the Asian continent to the northwest of Japan provides favourable conditions for intense precipitation in the Baiu season in Japan (Kanada et al., 2010a). Kitoh et al. (2009) projected changes in precipitation characteristics during the east Asian summer rainy season, using a 5-km mesh cloud-resolving model embedded in a 20-km mesh global atmospheric model with CMIP3 mean SST changes. The frequency of heavy precipitation is projected to increase at the end of the 21st century for hourly as well as daily precipitation. Further, extreme hourly precipitation is projected to increase even in the near future (2030s) when the temperature increase is still modest, even though uncertainties in the projection (and even the simulation) of hourly rainfall are still high.

Climate change scenarios for the 21st century show a weakening of the North American monsoon through a weakening and poleward expansion of the Hadley cell (Lu et al., 2007). The expansion of the Hadley cell is caused by an increase in the subtropical static stability, which pushes poleward the baroclinic instability zone and hence the outer boundary of the Hadley cell. Simple physical arguments (Held and Soden, 2006) predict a slowdown of the tropical overturning circulation under global warming. A few studies (e.g., Marengo et al., 2009a) have projected over the period 1960-2100 a weak tendency for an increase of dry spells. The projections show an increase in the frequency of rainfall extremes in southeastern South America by the end of the 21st century, possibly due to an intensification of the moisture transport from Amazonia by a more frequent/intense low-level jet east of the Andes in the A2 emissions scenario (Marengo et al., 2009a; Soares and Marengo, 2009).

There are many deficiencies in model representation of the monsoons and the processes affecting them, and this reduces confidence in their ability to project future changes. Some of the uncertainty on global and regional climate change projections in the monsoon regions results from the model representation of resolved processes (e.g., moisture advection), the parameterizations of sub-grid-scale processes (e.g., clouds, precipitation), and model simulations of feedback mechanisms on the global and regional scale (e.g., changes in land-use/cover, see also Section 3.1.4). Kharin and Zwiers (2007) made an intercomparison of precipitation extremes in the tropical region in all AR4 models with observed extremes expressed as 20 year return values. They found a very large disagreement in the Tropics suggesting that some physical processes associated with extreme precipitation are not well represented by the models. Shukla (2007) noted that current climate models cannot even adequately predict the mean intensity and the seasonal variations of the Asian summer monsoon. This reduces confidence in the projected changes in extreme precipitation over the monsoon regions. Many of the important climatic effects of the Madden Julian Oscillation (MJO), including its impacts on rainfall variability in the monsoons, are still poorly simulated by contemporary climate models (Christensen et al., 2007). Current GCMs still have difficulties and display a wide range of skill in simulating the subseasonal variability associated with Asian summer monsoon (Lin et al., 2008b). Most GCMs simulate westward propagation of the coupled equatorial easterly waves, but relatively poor eastward propagation of the MJO and overly weak variances for both the easterly waves and the MJO. Most GCMs are able to reproduce the basic characteristics of the precipitation seasonal cycle associated with the South American Monsoon System (SAMS), but there are large discrepancies in the South Atlantic Convergence Zone represented by the models in both intensity and location, and in its seasonal evolution (Vera et al., 2006). In addition, models exhibit large discrepancies in the direction of the changes associated with the summer (SAMS) precipitation, which makes the projections for that tropical region highly uncertain. Lin et al. (2008a) show that the coupled GCMs have significant problems and display a wide range of skill in simulating the North American monsoon and associated intraseasonal variability. Most of the models reproduce the monsoon rain belt, extending from southeast to northwest, and its gradual northward shift in early summer, but overestimate the precipitation over the core monsoon region throughout the seasonal cycle and fail to reproduce the monsoon retreat in the fall. The AR4 assessed that models fail in representing the main features of the west African monsoon although most of them do have a monsoonal climate albeit with some distortion (Christensen et al., 2007). Other major sources of uncertainty in projections of monsoon changes are the responses and feedbacks of the climate system to emissions as represented in climate models. These uncertainties are particularly related to the representation of the conversion of the emissions into concentrations of radiatively active species (i.e., via atmospheric chemistry and carbon-cycle models) and especially those derived from aerosol products of biomass burning. The subsequent response of the physical climate system complicates the nature of future projections of monsoon precipitation. Moreover, the long-term variations of model skill in simulating monsoons and their variations represent an additional source of uncertainty for the monsoon regions, and indicate that the regional reliability of long climate model runs may depend on the time slice for which the output of the model is analyzed.

The AR4 (Hegerl et al., 2007) concluded that the current understanding of climate change in the monsoon regions remains one of considerable uncertainty with respect to circulation and precipitation. With few exceptions in some monsoon regions, this has not changed since. Since the above mentioned conclusions have been based on very few studies, and there are many issues with model representation of monsoons and the underlying processes, there is *low confidence* in projections of changes in monsoons, even in the sign of the change. However, one common pattern may be an increase in extreme precipitation (see 3.3.2), though not necessarily induced by changes in monsoon characteristics, and not necessarily in all monsoon regions.

3.4.2. El Niño – Southern Oscillation

The El Niño – Southern Oscillation (ENSO) is a natural fluctuation of the global climate system caused by equatorial ocean-atmosphere interaction in the tropical Pacific Ocean (Philander, 1990). The term "Southern Oscillation" refers to a tendency for above average surface atmospheric pressures in the Indian Ocean to be associated with below average pressures in the Pacific, and vice versa. This oscillation is associated with variations in sea surface temperatures (SST) in the east equatorial Pacific. The oceanic and atmospheric variations are collectively referred to as ENSO. An El Niño episode is one phase of the ENSO phenomenon and is associated with abnormally warm central and east equatorial Pacific Ocean surface temperatures, while the opposite phase, a La Niña episode, is associated with abnormally cool ocean temperatures in this region. Both extremes are associated with a characteristic spatial pattern of droughts and floods. An El Niño episode is usually accompanied by drought in southeastern Asia, India, Australia, southeastern Africa, Amazonia, and northeast Brazil, with fewer than normal tropical cyclones around Australia and in the North Atlantic. Wetter than normal conditions during El Niño episodes are observed along the west coast of tropical South America, subtropical latitudes of western North America and southeastern America. In a La Niña episode the climate anomalies are usually the opposite of those in an El Niño. Pacific islands are strongly affected by ENSO variations. Recent research (e.g., Kenyon and Hegerl, 2008; Ropelewski and Bell, 2008; Schubert et al., 2008a; Alexander et al., 2009; Grimm and Tedeschi, 2009; Zhang et al., 2010) has demonstrated that different phases of ENSO (El Niño or La Niña episodes) also are associated with different frequencies of occurrence of short-term weather extremes such as heavy rainfall events and extreme temperatures. The relationship between ENSO and interannual variations in tropical cyclone activity is well-known (e.g., Kuleshov et al., 2008). The simultaneous occurrence of a variety of climate extremes in an El Niño episode (or a La Niña episode) may provide special challenges for organizations coping with disasters induced by ENSO.

The AR4 noted that orbital variations could affect the ENSO behaviour (Jansen et al., 2007). Cane (2005) found that a relatively simple coupled model suggested that systematic changes in the El Niño could be stimulated by seasonal changes in solar insolation. However, a more comprehensive model simulation (Wittenberg, 2009) has suggested that long-term changes in the behaviour of the phenomenon might occur even without forcing from radiative changes. Vecchi and Wittenberg (2010) concluded that the "tropical Pacific could generate variations in ENSO frequency and intensity on its own (via chaotic behaviour), respond to external radiative forcings (e.g., changes in greenhouse gases, volcanic eruptions, atmospheric aerosols, etc), or both". Meehl et al. (2009a) demonstrate that solar insolation variations related to the 11-year sunspot cycle can affect ocean temperatures associated with ENSO.

ENSO has varied in strength over the last millennium with stronger activity in the 17th century and late 14th century, and weaker activity during the 12th and 15th centuries (Cobb et al., 2003; Conroy et al., 2009). On longer timescales, there is evidence that ENSO may have changed in response to changes in the orbit of the Earth (Vecchi and Wittenberg, 2010), with the phenomenon apparently being weaker around 6,000 years ago (according to proxy measurements from corals and climate model simulations) (Rein et al., 2005; Brown et al., 2006; Otto-Bliesner et al., 2009) and model simulations suggest that it was stronger at the Last Glacial Maximum or LGM (An et al., 2004). Fossil coral evidence indicates that the phenomenon did continue to operate during the LGM (Tudhope et al., 2001). Thus the paleoevidence indicates that ENSO can continue to operate, although altered perhaps in intensity, through quite anomalous climate periods.

The AR4 noted that the nature of ENSO has varied substantially over the period of instrumental data, with strong events from the late 19th century through the first quarter of the 20th century and again after 1950. An apparent climate shift around 1976–1977 was associated with a shift to generally above-normal SSTs in the central and eastern Pacific and a tendency towards more prolonged and stronger El Niño episodes (Trenberth et al., 2007). Ocean temperatures in the central equatorial Pacific (the so-called NINO3 index) suggest a trend toward more frequent or stronger El Niño episodes over the past 50–100 years (Vecchi and Wittenberg, 2010). Vecchi et al. (2006) reported a weakening of the equatorial Pacific pressure gradient since the 1960s, with a sharp drop in the 1970s. Power and Smith (2007) proposed that the apparent dominance of El Niño during the last few decades was due in part to a change in the background state of the Southern Oscillation Index or SOI (the standardized difference in surface atmospheric pressure between Tahiti and Darwin), rather than a change in variability or a shift to more frequent El Niño events alone. Nicholls (2008) examined the behaviour of the SOI and another index, the NINO3.4 index of central equatorial Pacific SSTs, but found no evidence of trends in the variability or the persistence of the indices, (although Yu and Kao (2007) reported decadal variations in the persistence barrier, the tendency for weaker persistence across the Northern Hemisphere spring), nor in their seasonal patterns. There was a trend towards what might be considered more ‘‘El Niño-like’’ behaviour in the SOI (and more weakly in NINO3.4), but only through the period March–September and not in November–February, the season when El Niño and La Niña events typically peak. The trend in the SOI reflected only a trend in Darwin pressures, with no trend in Tahiti pressures. Apart from this trend, the temporal/seasonal nature of the El Niño–Southern Oscillation has been remarkably consistent through a period of strong global warming. There is evidence, however, of a tendency for recent El Niño episodes to be centered more in the central equatorial Pacific than in the east Pacific (Yeh et al., 2009). In turn, this change in the location of the strongest SST anomalies associated with El Niño may explain changes that have been noted in the remote influences of the phenomenon on the climate over Australia and in the mid-latitudes (Wang and Hendon, 2007; Weng et al., 2009). For instance, Taschetto et al. (2009) demonstrated that episodes with the warming centred in the central Pacific exhibit different patterns of Australian rainfall variations relative to the east Pacific centred El Niño events.

The possible role of increased greenhouse gases in affecting the behaviour of ENSO over the past 50-100 years is uncertain. Yeh et al. (2009) suggested that changes in the background temperature associated with increases in greenhouse gases should affect the behaviour of the El Niño, such as the location of the strongest SST anomalies, because El Niño behaviour is strongly related to the average ocean temperature gradients in the equatorial Pacific. Some studies (e.g., Zhang et al., 2008a) have suggested that increased activity might be due to increased CO_2, however no formal attribution study has yet been completed and some other studies (e.g., Power and Smith, 2007) suggest that changes in the phenomenon are within the range of natural variability (i.e., that no change has yet been detected, let alone attributed).

Global warming is projected to lead to a mean reduction of the zonal winds across the equatorial Pacific (Vecchi and Soden, 2007b). However, this change should not be described as an “El Niño – like” average change even though during an El Niño episode these winds also weaken, because there is only limited correspondence between these changes in mean state of the equatorial Pacific and an El Niño episode. AR4 determined that all models exhibited continued ENSO interannual variability in projections through the 21st century, but the projected behaviour of the phenomenon differed between models, and it was concluded that “there is no consistent indication at this time of discernible changes in projected ENSO amplitude or frequency in the 21st century” (Meehl et al., 2007b). Models project a wide variety of changes in ENSO variability and the frequency of El Niño episodes as a consequence of increased greenhouse gas concentrations, with a range between a 30% reduction to a 30% increase in variability (van Oldenborgh et al., 2005). One model study even found that although ENSO activity increased when CO_2 concentrations were doubled or quadrupled, a considerable decrease in activity occurred when CO_2 was increased by a factor of 16 times, much greater than is possible through the 21st century (Cherchi et al., 2008), suggesting a high variability of possible ENSO changes as a result of CO_2 changes. The remote impacts, on rainfall for instance, of ENSO may change as CO_2 increases, even if the equatorial Pacific aspect of ENSO does not change substantially. For instance, regions in which rainfall increases in the future tend to show increases in interannual rainfall variability (Boer, 2009), without any strong change in the interannual variability of tropical SSTs. Also, since some long-term projected changes in response to increased greenhouse gases may resemble the climate response to an El Niño event, this may enhance or mask the response to El Niño events in the future (Lau et al., 2008b; Müller and Roeckner, 2008).

One change that models tend to project is an increasing tendency for El Niño episodes to be centred in the central equatorial Pacific, rather than the traditional location in the eastern equatorial Pacific. Yeh et al. (2009) examined the relative frequency of El Niño episodes simulated in coupled climate models with projected increases in greenhouse gas concentrations. A majority of models, especially those best able to simulate the current ratio of central Pacific locations to east Pacific locations of El Niño events, projected a further increase in the relative frequency of these central Pacific events. Such a change would also have implications for the remote influence of the phenomenon on climate away from the equatorial Pacific (e.g., Australia and India). However, even the projection that the 21st century may see an increased frequency of central Pacific El Niño episodes, relative to the frequency of events located further east (Yeh et al., 2009), is subject to considerable uncertainty. Of the 11 coupled climate model simulations examined by Yeh et al. (2009), three projected a relative decrease in the frequency of these central Pacific episodes, and only four of the models produced a statistically significant change to more frequent central Pacific events.

A caveat regarding all projections of future behaviour of ENSO arises from systematic biases in the depiction of the ENSO behaviour through the 20th century by models (Randall et al., 2007; Guilyardi et al., 2009). Leloup et al. (2008) for instance, demonstrate that coupled climate models show wide differences in the ability to reproduce the spatial characteristics of SST variations associated with ENSO during the 20th century, and all models have failings. They concluded that it is difficult to even classify models by the quality of their reproductions of the behaviour of ENSO, because models scored unevenly in their reproduction of the different phases of the phenomenon. This makes it difficult to determine which models to use to project future changes of the ENSO. Moreover, most of the models are not able to reproduce the typical wavetrains observed in the circulation anomalies associated with ENSO in the Southern Hemisphere (Vera and Silvestri, 2009) and the Northern Hemisphere (Joseph and Nigam, 2006).

The position at the time of the AR4 was that there was no consistency of projections of changes in ENSO variability or frequency in the future (Meehl et al., 2007b). This position has not been changed as a result of post-AR4 studies. The evidence is that the nature of the ENSO has varied in the past apparently sometimes in response to changes in radiative forcing but also possibly due to internal climatic variability. Since radiative forcing will continue to change in the future, we can confidently expect changes in the ENSO will as well. However, Vecchi and Wittenberg (2010) conclude "the ENSO variations we see in decades to come may be different than those we've seen in recent decades – yet we are not currently at a state to confidently project what those changes will be". They also observe that El Niño and La Niña events will continue to occur and influence the climate but that there will continue to be variations in the phenomenon and its impacts, on a variety of timescales. Similarly, Collins et al. (2010) conclude that "despite considerable progress in our understanding of the impact of climate change on many of the processes that contribute to El Niño variability, it is not yet possible to say whether ENSO activity will be enhanced or damped, or if the frequency of events will change."

In summary, models project a wide variety of changes in El Niño – Southern Oscillation variability and the frequency of El Niño episodes as a consequence of increased greenhouse gas concentrations, and so there is *low confidence* in projections of changes in the phenomenon. However, there is *medium confidence* regarding a projected increase (simulated by most GCMs) in the relative frequency of central equatorial Pacific events, which typically exhibit different patterns of climate variations than do the classical East Pacific events.

3.4.3. Other Modes of Variability

Other natural modes of variability that are relevant to extremes and disasters include the North Atlantic Oscillation (NAO), the Southern Annular Mode (SAM) and the Indian Ocean Dipole (IOD; Trenberth et al., 2007). The NAO is a large-scale seesaw in atmospheric pressure between the subtropical high and the polar low in the North Atlantic region. The positive NAO phase has a strong subtropical high-pressure center and a deeper than normal Icelandic low. This results in a shift of winter storms crossing the Atlantic Ocean to a more northerly track, and is associated with warm and wet winters in northwestern Europe and cold and dry winters in northern Canada and Greenland. Scaife et al. (2008) discuss the relationship between the NAO and European extremes. The NAO is closely related to the Northern Annular Mode (NAM); for brevity we focus here on the NAO but much of what is said about the NAO also applies to the NAM. The SAM refers to north-south shifts in atmospheric mass between the Southern Hemisphere middle and high latitudes and is the most important pattern of climate variability in these latitudes. The SAM positive phase is linked to negative sea level pressure anomalies over the polar regions and intensified westerlies. It has been associated with cooler than normal temperatures over most of Antarctica and Australia, with warm anomalies over the Antarctic Peninsula, southern South America, and southern New Zealand, and with anomalously dry conditions over southern South America, New Zealand, and Tasmania and wet anomalies over much of Australia and South Africa (e.g., Hendon et al., 2007). The IOD is a coupled ocean-atmosphere phenomenon in the Indian Ocean. A positive IOD event is associated with anomalous cooling in the southeastern equatorial Indian Ocean and anomalous warming in the western equatorial Indian Ocean. Recent work (Ummenhofer et al., 2008; 2009a; 2009b) has implicated the IOD as a cause of droughts in Australia, and heavy rainfall in east Africa (Ummenhofer et al., 2009c). There is also evidence of modes of variability operating on multi-decadal time-scales, notably the Pacific Decadal Oscillation (PDO) and the Atlantic

Multi-decadal Oscillation (AMO). Variations in the PDO have been related to precipitation extremes over North America (Zhang et al., 2010).

Both the NAO and the SAM have exhibited trends towards their positive phase (strengthened mid-latitude westerlies) over the last three to four decades, although both have returned to near their long-term mean state in the last five years (Trenberth et al., 2007). The AR4 (Hegerl et al., 2007) noted that trends over recent decades in the NAO and SAM are *likely* related in part to human activity. Goodkin et al. (2008) conclude that the variability in the NAO is linked with changes in the mean temperature of the Northern Hemisphere. Dong et al. (2010) demonstrated that some of the observed late 20th century decadal-scale changes in NAO behaviour could be reproduced by increasing the CO_2 concentrations in a coupled model, and concluded that greenhouse gas concentrations may have played a role in forcing these changes. The largest observed trends in SAM occur in December-February, and model simulations indicate that these are due mainly to ozone changes. However it has been argued that anthropogenic circulation changes are poorly characterized by trends in the annular modes (Woollings et al., 2008). Further complicating these trends, Silvestri and Vera (2009) reported changes in the typical hemispheric circulation pattern related to SAM and its associated impact on both temperature and precipitation anomalies, particularly over South America and Australia, between the 1960s–70s and 1980s–90s. The time scales of variability in modes such as the AMO and PDO are so long that it is difficult to diagnose any change in their behaviour in modern data.

The AR4 noted that there was considerable spread among the model projections of the NAO, leading to low confidence in NAO projected changes, but the magnitude of the increase for the SAM is generally more consistent across models (Meehl et al., 2007b). However, the ability of coupled models to simulate the observed SAM impact on climate variability in the Southern Hemisphere is limited (e.g., Miller et al., 2006; Vera and Silvestri, 2009). Variations in the longer time-scale modes of variability (AMO, PDO) might affect projections of changes in extremes associated with the various natural modes of variability and global temperatures (Keenlyside et al., 2008).

Sea level pressure is projected to increase over the subtropics and mid-latitudes, and decrease over high latitudes (Meehl et al., 2007b). This would equate to trends in the NAO and SAM, with a poleward shift of the storm tracks of several degrees latitude and a consequent increase in cyclonic circulation patterns over the Arctic and Antarctica. During the 21st century, although stratospheric ozone concentrations are expected to recover, tending to lead to a weakening of the SAM, polar vortex intensification is likely to continue due to the increases in greenhouse gases. A recent study (Woolings et al., 2010) found a tendency towards a more positive NAO under anthropogenic forcing through the 21st century, although they concluded that confidence in the model projections was low because of deficiencies in its simulation of current-day NAO regimes. Goodkin et al. (2008) predict continuing high variability, on multidecadal scales, in the NAO with continued global warming. Keenlyside et al. (2008) proposed that variations associated with the multi-decadal modes of variability may offset warming due to increased greenhouse gas concentrations over the next decade or so. Conway et al. (2007) reported that model projections of future IOD behaviour showed no consistency. Kay and Washington (2008) reported that under some emissions scenarios, changes in a dipole mode in the Indian Ocean could change rainfall extremes in southern Africa.

In summary, issues with the ability of models to simulate current behaviour of these natural modes, the likely influence of competing factors (e.g., ozone, greenhouse gases) on current and future mode behaviour, and inconsistency between the model projections, means that there is *low confidence* in the ability to project changes in the modes.

3.4.4. Tropical Cyclones

Tropical cyclones occur in most tropical oceans and pose a significant threat to coastal populations and infrastructure, and marine interests such as shipping and offshore activities. Each year, about 90 tropical cyclones occur globally, and this number has remained roughly steady over the modern period of geostationary satellites (since around the mid-1970s). While the global frequency has remained steady, there can be substantial inter-annual to multi-decadal frequency variability within individual ocean basins (e.g., Webster et al., 2005). This regional variability, particularly when combined with substantial inter-annual to multi-decadal variability in tropical cyclone tracks (e.g., Kossin et al., 2010), presents a significant challenge for disaster planning and mitigation aimed at specific regions.

Tropical cyclones are perhaps most commonly associated with extreme wind, but storm-surge and fresh-water flooding from extreme rainfall generally cause the great majority of damage and loss of life (e.g., Rappaport, 2000; Webster, 2008). Related indirect factors, such as the failure of the levee system in New Orleans during the passage of Hurricane Katrina (2005), or mudslides during the landfall of Hurricane Mitch (1998) in Central America, represent important related impacts. Projected sea level rise will further compound tropical cyclone surge impacts. Tropical cyclones that track northward (southward) in the Northern (Southern) hemisphere can undergo a transition to become extratropical cyclones. While these storms have different characteristics than their tropical progenitors, they can still be accompanied by a storm surge that can impact regions well away from the tropics (e.g., Danard et al., 2004).

Tropical cyclones are typically classified in terms of their intensity, which is a measure of near-surface wind speed (sometimes categorized according to the Saffir-Simpson scale). The strongest storms (Saffir-Simpson category 4 and 5) are comparatively rare but are generally responsible for the majority of damage (e.g., Landsea, 1993; Pielke et al., 2008). Additionally, there are marked differences in the characteristics of both observed and projected tropical cyclone variability when comparing weaker and stronger tropical cyclones (e.g., Webster et al., 2005; Elsner et al., 2008; Bender et al., 2010), while records of the strongest storms are potentially less reliable than those of their weaker counterparts (Landsea et al., 2006).

While there is a relationship between intensity and storm surge, the structure and areal extent of the wind field also play an important role. Other relevant tropical cyclone measures include frequency, duration, and track. Forming robust physical links between all of the metrics briefly mentioned here and natural or human-induced climate variability is a major challenge. Significant progress is being made, but substantial uncertainties still remain due largely to data quality issues (see 3.2.1, and below) and imperfect theoretical and modeling frameworks (see below).

Detection of trends in tropical cyclone metrics such as frequency, intensity, and duration remains a significant challenge. Historical tropical cyclone records are known to be heterogeneous due to changing observing technology and reporting protocols (e.g., Landsea et al., 2004). Further heterogeneity is introduced when records from multiple ocean basins are combined to explore global trends because data quality and reporting protocols vary substantially between regions (Knapp and Kruk, 2010). Progress has been made toward a more homogeneous global record of tropical cyclone intensity using satellite data (Knapp and Kossin, 2007; Kossin et al., 2007), but these records are necessarily constrained to the satellite era and so only represent the past 30-40 years.

Natural variability combined with uncertainties in the historical data makes it difficult to detect trends in tropical cyclone activity. There have been no significant trends observed in global tropical cyclone frequency records, including over the present 40-year period of satellite observations (e.g., Webster et al., 2005). Regional trends in tropical cyclone frequency have been identified in the North Atlantic, but the fidelity of these trends is debated (Holland and Webster, 2007; Landsea, 2007; Mann et al., 2007b). Landsea et al. (2010) showed that a large contribution of the observed long-term trend in the record of North Atlantic tropical cyclone frequency is due to a trend in the frequency of short-lived storms, a subset of storms that may be particularly sensitive to changes in technology and reporting protocols. However, Emanuel (2010) demonstrates that the changes in short-duration storms may also have physical causes, and Kossin et al. (2010) find that much of the changes in the frequency of short-duration storms in the Atlantic have occurred in the Gulf of Mexico in close proximity to land and thus largely avoids the data-quality issues with pre-satellite storm undercounts.

Different methods for estimating undercounts in the earlier part of the North Atlantic tropical cyclone record provide mixed conclusions (Chang and Guo, 2007; Mann et al., 2007a; Kunkel et al., 2008; Vecchi and Knutson, 2008). Regional trends have not been detected in other oceans (Chan and Xu, 2009; Kubota and Chan, 2009). It thus remains uncertain whether any reported long-term increases in tropical cyclone frequency are robust, after accounting for past changes in observing capabilities (Knutson et al., 2010).

Whereas frequency estimation requires only that a tropical cyclone be identified and reported at some point in its lifetime, intensity estimation requires a series of specifically targeted measurements over the entire duration of the tropical cyclone (e.g., Landsea et al., 2006). Consequently, intensity values in the historical records are especially sensitive to changing technology and improving methodology, which heightens the challenge of detecting trends within the backdrop of natural variability. Global reanalyses of tropical cyclone intensity using a homogenous satellite record have suggested that changing technology has introduced a non-stationary bias that inflates trends in measures of intensity (Kossin et al., 2007), but a significant upward trend in the intensity of the strongest tropical cyclones remains after this bias is accounted for (Elsner et al., 2008). While these analyses are suggestive of a link between observed tropical cyclone intensity and climate change, they are necessarily confined to a 30+ year period of satellite observations, and cannot provide clear evidence for a longer-term trend.

Time series of power dissipation, an aggregate compound of tropical cyclone frequency, duration, and intensity that measures total energy consumption by tropical cyclones, show upward trends in the North Atlantic and weaker upward trends in the western North Pacific over the past 25 years (Emanuel, 2007), but interpretation of longer-term trends is again constrained by data quality concerns. The variability and trend of power dissipation can be related to SST and other local factors such as tropopause temperature, and vertical wind shear (Emanuel, 2007), but it is a current topic of debate whether local SST or the difference between local SST and mean tropical SST is the more physically relevant metric (Swanson, 2008). The distinction is an important one when making projections of power dissipation based on projections of SST, particularly in the Atlantic where SST has been increasing more rapidly than the tropics as a whole (Vecchi et al., 2008). Since 2005, accumulated cyclone energy, which is an integrated metric analogous to power dissipation, has been declining globally and is presently at a 40-year low point (Maue, 2009). The present period of quiescence, as well as the period of heightened activity leading up to a high point in 2005, do not clearly represent substantial departures from past variability (Maue, 2009).

Increases in tropical water vapor and rainfall (Trenberth et al., 2005; Lau and Wu, 2007) have been identified and there is some evidence for related changes in tropical cyclone-related rainfall (Lau et al., 2008a), but a robust and consistent trend in tropical cyclone rainfall has not yet been established due to a general lack of studies. Similarly, increases in the length of the North Atlantic hurricane season have been noted (Kossin, 2008), but the uncertainty in the amplitude of the trends and the lack of additional studies limits the utility of these results for a meaningful assessment.

Estimates of tropical cyclone variability prior to the modern instrumental historical record have been constructed using archival documents (Chenoweth and Devine, 2008), coastal marsh sediment records and isotope markers in coral, speleothems, and tree-rings, among other methods (Frappier et al., 2007a). These estimates demonstrate centennial- to millennial-scale relationships between climate and tropical cyclone activity (Donnelly and Woodruff, 2007; Frappier et al., 2007b; Nott et al., 2007; Nyberg et al., 2007; Scileppi and Donnelly, 2007; Neu, 2008; Woodruff et al., 2008a; Woodruff et al., 2008b; Mann et al., 2009; Yu et al., 2009) but generally do not provide robust evidence that the observed post-industrial tropical cyclone activity is unprecedented.

The AR4 Summary for Policy Makers concluded that it is likely that a trend had occurred in intense tropical cyclone activity since 1970 in some regions (IPCC, 2007b). In somewhat more detail, it was further stated that "*there is observational evidence for an increase in intense tropical cyclone activity in the North Atlantic since about 1970, correlated with increases of tropical SSTs. There are also suggestions of increased intense tropical cyclone activity in some other regions where concerns over data quality are greater. Multi-decadal variability and the quality of the tropical cyclone records prior to routine satellite observations in about 1970 complicate the detection of long-term trends in tropical cyclone activity. There is no clear trend in the annual numbers of tropical cyclones*."

The subsequent U.S. CCSP SAP 3.3 (Kunkel et al., 2008) concluded that "*Atlantic tropical storm and hurricane destructive potential as measured by the Power Dissipation Index (which combines storm intensity, duration, and frequency) has increased".* The report concluded that *"…the power dissipation increase is substantial since about 1970, and is likely substantial since the 1950s and 60s, in association with warming Atlantic SSTs*", and that "*it is likely that the annual numbers of tropical storms, hurricanes and major hurricanes in the North Atlantic have increased over the past 100 years, a time in which Atlantic SSTs also increased*", but that "*the evidence is not compelling for significant trends beginning in the late 1800s*". Based on research subsequent to the IPCC AR4 and CCSP SAP3.3, which further elucidated the scope of uncertainties in the historical tropical cyclone data, the most recent assessment by the World Meteorological Organization Expert Team on Climate Change Impacts on Tropical Cyclones (Knutson et al., 2010) does not conclude that it is likely that annual numbers of tropical storms, hurricanes and major hurricanes counts have increased over the past 100 years in the North Atlantic basin, nor does it conclude that the Atlantic Power Dissipation Index increase is "likely substantial" since the 1950s and 60s.

The present assessment regarding observed trends in tropical cyclone activity is essentially unchanged from the AR4 and the WMO report (Knutson et al., 2010), but differs from the CCSP SAP 3.3 report based on subsequent research elucidating the scope of uncertainties in the historical tropical cyclone data: There is *low confidence* that any reported long-term increases in tropical cyclone activity are robust, after accounting for past changes in observing capabilities.

In addition to the natural variability of tropical SSTs, several studies have concluded that there is a detectable tropical SST warming trend due to increasing greenhouse gases (Karoly and Wu, 2005; Knutson et al., 2006; Santer et al., 2006; Gillett et al., 2008a). The region where this anthropogenic warming has occurred encompasses tropical cyclogenesis regions, and the CCSP SAP 3.3 report (2008) stated that “it is very likely that human-caused increases in greenhouse gases have contributed to the increase in SSTs in the North Atlantic and the Northwest Pacific hurricane formation regions over the 20th century.” Changes in the mean thermodynamic state of the tropics can be directly linked to tropical cyclone variability within the theoretical framework of potential intensity theory (Bister and Emanuel, 1998). In this framework, the expected response of tropical cyclone intensity to observed climate change is relatively straightforward: if climate change causes an increase in the ambient potential intensity that tropical cyclones move through, the distribution of intensities in a representative sample of storms is expected to shift toward greater intensities (Emanuel, 2000; Wing et al., 2007). Such a shift in the distribution would be most evident at the upper quantiles of the distribution as the strongest tropical cyclones become stronger (Elsner et al., 2008).

Given the evidence that SST in the tropics has increased due to increasing greenhouse gases, and the theoretical expectation that increases in potential intensity will lead to stronger storms, it is essential to fully understand the relationship between SST and potential intensity. Observations demonstrate a strong positive correlation between SST and the potential intensity. This relationship suggests that SST increases will lead to increased potential intensity, which will then ultimately lead to stronger storms (Emanuel, 2000; Wing et al., 2007). However, there is a growing body of research suggesting that local potential intensity is controlled by the difference between local SST and spatially averaged SST in the tropics (Vecchi and Soden, 2007a; Xie et al., 2010; Ramsay and Sobel, 2011). Since increases of SST due to global warming are not expected to lead to continuously increasing SST gradients, this recent research suggests that increasing SST due to global warming, by itself, does not yet have a fully-understood physical link to increasingly strong tropical cyclones.

The present period of heightened tropical cyclone activity in the North Atlantic, concurrent with comparative quiescence in other ocean basins (e.g., Maue, 2009), is apparently related to differences in the rate of SST increases, as global SST has been rising steadily but at a slower rate than the Atlantic (Holland and Webster, 2007). The present period of relatively enhanced warming in the Atlantic has been proposed to be due to internal variability (Zhang and Delworth, 2009), anthropogenic tropospheric aerosols (Mann and Emanuel, 2006), and mineral (dust) aerosols (Evan et al., 2009). None of these proposed mechanisms provide a clear expectation that North Atlantic SST will continue to increase at a greater rate than the tropical-mean SST.

Changes in tropical cyclone intensity, frequency, genesis location, duration, and track contribute to what is sometimes broadly defined as "tropical cyclone activity". Of these metrics, intensity has the most direct physically reconcilable link to climate variability within the framework of potential intensity theory, as described above (Kossin and Vimont, 2007). Statistical correlations between necessary ambient environmental conditions (e.g., low vertical wind shear and adequate atmospheric instability and moisture) and tropical cyclogenesis frequency have been well documented (DeMaria et al., 2001) but changes in these conditions due specifically to increasing greenhouse gas do not necessarily preserve the same statistical relationships. For example, the observed minimum SST threshold for tropical cyclogenesis is roughly 26°C. This relationship might lead to an expectation that anthropogenic warming of tropical SST and the resulting increase in the areal extent of the region of 26°C SST should lead to increases in tropical cyclone frequency. However, there is a growing body of evidence that the minimum SST threshold for tropical cyclogenesis increases at about the same rate as the SST increase due solely to CO_2 forcing (e.g., Ryan et al., 1992; Dutton et al., 2000; Yoshimura et al., 2006; Bengtsson et al., 2007; Knutson et al., 2008; Johnson and Xie, 2010). That is, when the SST changes due to greenhouse warming are deconvolved from the background natural variability, that part of the SST variability may have no clear effect on tropical cyclogenesis. In this case, the simple observed relationship between tropical cyclogenesis and SST, while robust, does not adequately capture the relevant physical mechanisms of tropical cyclogenesis.

Another challenge to identifying causes behind observed changes in tropical cyclone activity is introduced by uncertainties in the reanalysis data used to identify environmental changes in regions where tropical cyclones develop and evolve (Bister and Emanuel, 2002; Emanuel, 2010). In particular, heterogeneity in upper-tropospheric kinematic and thermodynamic metrics complicate the interpretation of long-term changes in vertical wind shear and potential intensity, both of which are important environmental controls of tropical cyclones.

Based on a variety of model simulations, the expected long-term changes in tropical cyclone characteristics under greenhouse warming is a decrease or little change in frequency concurrent with an increase in mean intensity. One of the challenges for identifying these changes in the existing data records is that the expected changes predicted by the models are generally small when compared with changes associated with observed short-term natural variability. Based on changes in tropical cyclone intensity predicted by idealized numerical simulations with CO_2-induced tropical SST warming, Knutson and Tuleya (2004) suggested that clearly detectable increases may not be manifest for decades to come. Their argument was based on an informal comparison of the amplitude of the modelled upward trend (i.e., the signal) in storm intensity with the amplitude of the interannual variability (i.e., the noise). The recent high-resolution dynamical downscaling study of Bender et al. (2010) supports this argument and suggests that the predicted increases in the frequency of the strongest Atlantic storms may not emerge as a clear statistically significant signal until the latter half of the 21st century under the SRES A1B warming scenario.

With the exception of the North Atlantic, much of the global tropical cyclone data is confined to the period from the mid-20th century to present. In addition to the limited period of record, the uncertainties in the historical tropical cyclone data (Section 3.2.1 and above) and the extent of tropical cyclone variability due to random processes and linkages with various climate modes such as El Niño, do not presently allow for the detection of any clear trends in tropical cyclone activity that can be attributed to greenhouse warming. As such, it remains unclear to what degree the causal phenomena described here have modulated post-industrial tropical cyclone activity.

The AR4 concluded that "it is more likely than not that anthropogenic influence has contributed to increases in the frequency of the most intense tropical cyclones" (Hegerl et al., 2007). Based on subsequent research that further elucidated the scope of uncertainties in both the historical tropical cyclone data and the physical mechanisms underpinning the observed relationships, no such attribution conclusion was drawn in the recent WMO report (Knutson et al., 2010), which states on p. 14 of their Supplementary Information "…*we do not draw such an attribution conclusion in this assessment. Specifically we do not conclude that there has been a detectable change in tropical cyclone metrics relative to expected variability from natural causes, particularly owing to concerns about limitations of available observations and limited understanding of the possible role of natural climate variability in producing low frequency changes in the tropical cyclone metrics examined.*"

The present assessment regarding detection and attribution of trends in tropical cyclone activity is essentially unchanged from the WMO report (Knutson et al., 2010). The uncertainties in the historical tropical cyclone records, the incomplete understanding of the physical mechanisms linking tropical cyclone metrics to climate change, and the degree of tropical cyclone variability — comprising random processes and linkages to various natural climate modes such as El Niño — provide only low confidence for the attribution of any detectable changes in tropical cyclone activity to anthropogenic influences.

The AR4 concluded (Meehl et al., 2007b) that "*results from embedded high-resolution models and global models, ranging in grid spacing from 100 km to 9 km, project a likely increase of peak wind intensities and notably, where analysed, increased near-storm precipitation in future tropical cyclones. Most recent published modelling studies investigating tropical storm frequency simulate a decrease in the overall number of storms, though there is less confidence in these projections and in the projected decrease of relatively weak storms in most basins, with an increase in the numbers of the most intense tropical cyclones.*" The conclusions here are similar to those in the AR4, but somewhat more detail is now possible.

The spatial resolution of models such as the CMIP3 coupled ocean-atmosphere models used in the AR4 is generally not high enough to accurately resolve tropical cyclones, and especially to simulate their intensity (Randall et al., 2007). Higher resolution global models have had some success in reproducing tropical cyclone-like vortices (e.g., Chauvin et al., 2006; Oouchi et al., 2006; Zhao et al., 2009), but only their coarse characteristics. Significant progress has been recently made, however, using downscaling techniques whereby high-resolution models capable of reproducing more realistic tropical cyclones are run using boundary conditions provided by either reanalysis data sets or output fields from lower resolution climate models such as those used in the AR4 (e.g., Knutson et al., 2007; Emanuel et al., 2008; Knutson et al., 2008; Emanuel, 2010). A recent study by Bender et al. (2010) applies a cascading technique that downscales first from global to regional scale, and then uses the simulated storms from the regional model to initialize a very high resolution hurricane forecasting model. These downscaling studies have been increasingly successful at reproducing observed tropical cyclone characteristics, which provides increased confidence in their projections, and it is expected that more progress will be made as computing resources improve.

While detection of long-term past increases in tropical cyclone activity is complicated by data quality and signal-over-noise issues (as stated above), theory (Emanuel, 1987) and idealized dynamical models (Knutson and Tuleya, 2004) both predict increases in tropical cyclone intensity under greenhouse warming. The recent simulations with high-resolution dynamical models (Oouchi et al., 2006; Bengtsson et al., 2007; Gualdi et al., 2008; Knutson et al., 2008; Sugi et al., 2009; Bender et al., 2010) and statistical-dynamical models (Emanuel, 2007) consistently find that greenhouse warming causes tropical cyclone intensity to shift toward stronger storms by the end of the 21st century. These models also consistently project little change or a reduction in overall tropical cyclone frequency, but with an accompanying substantial fractional increase in the frequency of the strongest storms and increased precipitation rates in the models for which these metrics were examined. Mean 21st century global cyclone intensity changes under conditions roughly equivalent to A1B emissions scenarios are projected between 2 and 11%, and a decrease of –6 to –34% is projected in global tropical cyclone frequency. The downscaling experiments of Bender et al. (2010), which, as described above, use an ensemble of CMIP3 simulations to nudge a high-resolution dynamical model (Knutson et al., 2008) that is then used to initialize a very high-resolution dynamical model, project a 28% reduction in the overall frequency of Atlantic storms and an 80% increase in the frequency of Saffir-Simpson category 4 and 5 hurricanes over the next 80 years (A1B scenario). In addition to a decrease in frequency and an increase in intensity, higher resolution models also consistently project increased precipitation rates (~20%) within 100 km of storm centers.

The projected decreases in global tropical cyclone frequency may be due to increases in vertical wind shear (Vecchi and Soden, 2007c; Zhao et al., 2009; Bender et al., 2010), a weakening of the tropical circulation (Sugi et al., 2002; Bengtsson et al., 2007) associated with a decrease in the upward mass flux accompanying deep convection (Held and Soden, 2006), or an increase in the saturation deficit of the middle troposphere (Emanuel et al., 2008). For individual basins, there is much more uncertainty in projections of tropical cyclone frequency, with changes of up to ±50% or more projected by various models (Knutson et al., 2010). When projected SST changes are considered in the absence of projected radiative forcing changes, Northern Hemisphere tropical cyclone frequency has been found to increase (Wehner et al., 2010), which is congruent with the hypothesis that SST changes alone may not capture the relevant physical mechanisms controlling tropical cyclogenesis (e.g., Emanuel, 2010).

Another type of projection that is sometimes inferred from the literature is based on extrapolation of an observed statistical relationship (see also section 3.2.3). These relationships are typically constructed on past observed variability that represents a convolution of anthropogenically forced variability and natural variability across a broad range of timescales. In general however, these relationships cannot be expected to represent all of the relevant physics that control the phenomena of interest, and their extrapolation beyond the range of the observed variability they are built on is not reliable. As an example, there is a strong observed correlation between local SST and tropical cyclone power dissipation (Emanuel, 2007). If 21st century SST projections are applied to this relationship, power dissipation is projected to increase by about 300% in the next century (Vecchi et al., 2008; Knutson et al., 2010). Alternatively, there is a similarly strong relationship between power dissipation and relative SST, which represents the difference between local and tropical-mean SST and has been argued to serve as a proxy for local potential intensity (Vecchi and Soden, 2007a). When 21st century projections of relative SST are considered, this latter relationship projects almost no change of power dissipation in the next century (Vecchi et al., 2006). Both of these statistical relationships can be reasonably defended based on physical arguments but it is not clear which, if either, is correct (Ramsay and Sobel, 2011).

While projections under 21st century greenhouse warming indicate that it is *likely* that the global frequency of tropical cyclones will either decrease or remain essentially unchanged, an increase in mean tropical cyclone maximum wind speed (+2 to +11% globally) is also likely, although increases may not occur in all tropical regions (Knutson et al., 2010). Furthermore, while it is likely that overall global frequency will either decrease or remain essentially unchanged, there is *medium confidence* that the frequency of the most intense storms (e.g., Saffir-Simpson Category 4-5) will increase in some ocean basins. As noted above, observed changes in tropical cyclone-related rainfall have not been clearly established. However, as water vapour in the tropics increases (Trenberth et al., 2005) there is an expectation for increased tropical cyclone-related rainfall in response to associated moisture convergence increases (Held and Soden, 2006). This increase is expected to be compounded by increases in intensity as dynamical convergence under the storm is enhanced. Models in which tropical cyclone precipitation rates have been examined are highly consistent in projecting increased rainfall within the area near the tropical cyclone center under 21st century warming, with increases of +3% to +37% (Knutson et al., 2010). Typical projected increases are near +20%. Based on the level of consistency among models, and physical reasoning, it is *likely* that tropical cyclone-related rainfall rates will increase with greenhouse warming. Confidence in future projections in particular ocean basins is undermined by the inability of global models to reproduce accurate details at scales relevant to tropical cyclone genesis, track, and intensity evolution. Of particular concern is the limited ability of global models to accurately simulate upper-tropospheric wind (Cordero and Forster, 2006; Bender et al., 2010), which modulates vertical wind shear and tropical cyclone genesis and intensity evolution.

When simulating 21st century warming under the A1B emission scenario (or a close analogue), the present models and downscaling techniques as a whole are consistent in projecting (1) decreases or no change in tropical cyclone frequency, (2) increases in intensity and fractional increases in number of most intense storms, and (3) increases in tropical cyclone-related rainfall rates. Differences in regional projections lead to lower confidence in basin-specific projections of intensity, rainfall, and confidence is particularly low for projections of frequency within individual basins. Current models project frequency changes ranging from -6 to -34% globally, and up to ± 50% or more in individual basins by the late 21st century. There is low confidence in projections of changes in tropical cyclone genesis, location, tracks, duration, or areas of impact, and existing model projections do not show dramatic large-scale changes in these features.

In summary, there is *low confidence* that any reported long-term increases in tropical cyclone activity are robust, after accounting for past changes in observing capabilities. The uncertainties in the historical tropical cyclone records, the incomplete understanding of the physical mechanisms linking tropical cyclone metrics to climate change, and the degree of tropical cyclone variability provide only *low confidence* for the attribution of any detectable changes in tropical cyclone activity to anthropogenic influences. There is *low confidence* in projections of changes in tropical cyclone genesis, location, tracks, duration, or areas of impact. Based on the level of consistency among models, and physical reasoning, it is *likely* that tropical cyclone-related rainfall rates will increase with greenhouse warming. It is *likely* that the global frequency of tropical cyclones will either decrease or remain essentially unchanged, and an increase in mean tropical cyclone maximum wind speed (+2 to +11% globally) is *likely*, although increases may not occur in all tropical regions. While it is *likely* that overall global frequency will either decrease or remain essentially unchanged, there is *medium confidence* that the frequency of the most intense (e.g., Saffir-Simpson Category 4-5) storms will increase in some ocean basins.

3.4.5. Extratropical Cyclones

Extratropical cyclones (synoptic-scale low pressure systems) exist throughout the mid-latitudes in both hemispheres and mainly develop over the oceanic basins in the proximity of the upper tropospheric jet streams or as a result of flow over mountains (lee cyclogenesis). They are the main poleward transporter of heat and moisture and may be accompanied by adverse weather conditions such as windstorms, the build up of waves and storm surges or extreme precipitation events. Thus, changes in the intensity of extratropical cyclones or a systematic shift in the geographical location of extratropical cyclone activity may have a great impact on a wide range of regional climate extremes as well as the long-term changes in temperature and precipitation. Extratropical cyclones mainly form and grow via atmospheric instabilities such as a disturbance along a zone of strong temperature contrast (baroclinic instabilities), which is a reservoir of available potential energy that can be converted into the kinetic energy associated with extratropical cyclones. Intensification of the cyclones may also take place due to diabatic (temperature changes not related to adiabatic vertical displacement) processes such as release of energy due to phase changes of water (latent heat release) (Gutowski et al., 1992). Why should we expect climate change to influence extratropical cyclones? A simplified line of argument would be that both the low and high level pole to equator temperature gradients may change in a climate change scenario leading to a change in the atmospheric instabilities responsible for cyclone formation and growth (baroclinicity). Changes in precipitation intensities within extratropical cyclones will change the latent heat release. In addition, changes in the extratropical stormtracks are, according to theories on wave–mean flow interaction, associated with changes in the large-scale flow (Robinson, 2000; Lorenz and Hartmann, 2003). A latitudinal shift of the upper tropospheric jet would be accompanied by a latitudinal shift in the extratropical storm track. It is however still unclear to what extent a latitudinal shift of the jet changes the storm track activity rather than shifting it latitudinally (Wettstein and Wallace, 2010). Even within the simplified outline above the possible impact of climate change on the extratropical cyclone development is clearly not trivial.

When validated using reanalyses with similar horizontal resolution, modern climate models are found to represent the general structure of the storm track pattern well (Bengtsson et al., 2006; Greeves et al., 2007; Ulbrich et al., 2008; Catto et al., 2010). However, using data from five different coupled models the rate of transfer of zonal available potential energy to eddy available potential energy in synoptic systems was found to be too large yielding too much energy and an overactive energy cycle (Marques et al., 2011). Models tend to have too zonal stormtracks and some show a poor extension of the stormtracks into Europe (Pinto et al., 2006; Greeves et al., 2007; Orsolini and Sorteberg, 2009). It has also been noted that representation of cyclone activity may depend on the dynamical core and the horizontal resolution of the model (Jung et al., 2006; Greeves et al., 2007).

Paleoclimatic proxies for extratropical cyclone variability are still few, but progress is being made in using coastal dunefield development and sand grain content of peat bogs as proxies for storminess. Papers covering parts of western Europe indicate enhanced sand movement in European coastal areas during the Little Ice Age (Wilson et al., 2004; de Jong et al., 2006; Clemmensen et al., 2007; de Jong et al., 2007; Clarke and Rendell, 2009; Sjogren, 2009). It should be noted that sand influx is also influenced by sediment availability, which is controlled mainly by the degree of vegetation cover and the moisture content of the sediment (Li et al., 2004; Wiggs et al., 2004). Intense cultivation, overgrazing and forest disturbance make soils more prone to erosion, which can lead to increased sand transport even under less windy conditions.

Century-long time-series of estimates of extremes in geostrophic wind deduced from triangles of pressure stations, pressure tendencies from single stations (see section 3.3.3 for details) or oceanic variables such as extremes in non-tide residuals are (if these are located in the vicinity of the main stormtracks) possible proxies for extratropical cyclone activity. Trend detection in extratropical cyclone variables such as number of cyclones, intensity, and activity (parameters integrating cyclone intensity, number and possibly duration) became possible with the development of reanalyses, but remains challenging. Problems with the reanalyses have been especially pronounced in the southern hemisphere (Hodges et al., 2003; Wang et al., 2006a). Even though different reanalyses correspond well in the Northern Hemisphere (Hodges et al., 2003; Hanson et al., 2004) changes in the observing system giving artificial trends in integrated water vapor and kinetic energy (Bengtsson et al., 2004) may have influenced trends in both the number and intensity of cyclones. In addition, studies indicate that the magnitude and even the existence of the changes may depend on the choice of reanalysis (Simmonds et al., 2008) and cyclone tracking algorithm (Raible et al., 2008).

The AR4 noted a *likely* net increase in frequency/intensity of Northern Hemisphere extreme extratropical cyclones and a poleward shift in the tracks since the 1950s (Trenberth et al., 2007, Table 3.8), and cited several papers showing increases in the number or strength of intense extratropical cyclones both over the North Pacific and the North Atlantic storm track (Trenberth et al., 2007, p. 312), during the last 50 years. Studies using reanalyses indicate a northward and eastward shift in the Atlantic cyclone activity during the last 60 years with both more frequent and more intense wintertime cyclones in the high-latitude Atlantic (Weisse et al., 2005; Wang et al., 2006a; Schneidereit et al., 2007; Raible et al., 2008; Vilibic and Sepic, 2010) and fewer (Wang et al., 2006a; Raible et al., 2008) in the mid-latitude Atlantic. The increase in high-latitude cyclone activity was also reported in several studies of Arctic cyclone activity (Zhang et al., 2004c; Sorteberg and Walsh, 2008). Using ship-based trends in mean sea level pressure (MSLP) variance (which is tied to cyclone intensity), Chang (2007) found wintertime Atlantic trends to be consistent with (NCEP) reanalysis trends in the Atlantic, but slightly weaker. There are inconsistencies among studies of extreme cyclones in reanalyses, since some studies show an increase in intensity and number of extreme Atlantic cyclones (Geng and Sugi, 2001; Paciorek et al., 2002) while others show a reduction (Gulev et al., 2001). New studies have confirmed that positive (negative) NAM/NAO corresponds to stronger (weaker) Atlantic/European cyclone activity (e.g., Chang, 2009; Pinto et al., 2009). However, studies using long historical records also seem to suggest that some of these links may be intermittent (Hanna et al., 2008; Matulla et al., 2008; Allan et al., 2009) due to interdecadal shifts in the location of the positions of the NAO pressure centers (Vicente-Serrano and Lopez-Moreno, 2008; Zhang et al., 2008b). A possible influence of the PNA on the entrance of the North Atlantic stormtrack (over Newfoundland) has been reported by Pinto et al. (2011). It should be noted that there is some suggestion that the reanalyses cover a time period which starts with relatively low cyclonic activity in northern coastal Europe in the 1960s and reaches a maximum in the 1990s. Long-term European storminess proxies show no clear trends over the last century (see section 3.3.3 for details).

Studies using reanalyses and in situ data for the last 50 years have noted an increase in the number and intensity of north Pacific wintertime intense extratropical cyclone systems since the 1950s (Graham and Diaz, 2001; Simmonds and Keay, 2002; Raible et al., 2008) and cyclone activity (Zhang et al., 2004c), but signs of some of the trends disagreed when different tracking algorithms or reanalysis products are used (Raible et al., 2008). A slight positive trend has been found in north Pacific extreme cyclones (Geng and Sugi, 2001; Gulev et al., 2001; Paciorek et al., 2002). Using ship measurements, Chang (2007) found intensity related wintertime trends in the Pacific to be about 20%–60% of that found in the reanalysis. Long-term in situ observations of north Pacific cyclones based on observed pressure data are considerably fewer than for coastal Europe. However, using hourly tide gauge records at the western coast of the U.S. as a proxy for storminess, an increasing trend in the extreme winter non-tide residuals (NTR) has been observed the last decades (Bromirski et al., 2003; Menendez et al., 2008). Years having high NTR were linked to a large-scale atmospheric circulation pattern, with intense storminess associated with a broad, south-easterly displaced, deep Aleutian low that directed storm tracks toward the western U.S. coast. North Pacific cyclonic activity has been linked to tropical SST anomalies (NINO3.4) and PNA (Eichler and Higgins, 2006; Favre and Gershunov, 2006; Seierstad et al., 2007), showing that the PNA and NINO3.4 influence storminess, in particular over the eastern north Pacific with an equatorward shift in storm tracks in the North Pacific basin, as well as an increase of storm track activity along the U.S. East Coast during El Niño events.

Based on reanalyses North American cyclone numbers have increased over the last 50 years, with no statistically significant change in cyclone intensity (Zhang et al., 2004c). Hourly MSLP data from Canadian stations showed that winter cyclones have become significantly more frequent, longer lasting, and stronger in the lower Canadian Arctic over the last 50 years (1953–2002), but less frequent and weaker in the south, especially along the southeast and southwest Canadian coasts (Wang et al., 2006a). Further south, a tendency toward weaker low-pressure systems over the past few decades was found for U.S. east coast winter cyclones using reanalyses, but no statistically significant trends in the frequency of occurrence of systems (Hirsch et al., 2001).

Studies on extratropical cyclone activity in northern Asia are few. Using reanalyses a decrease in extratropical cyclone activity (Zhang et al., 2004c) and intensity (Zhang et al., 2004c; Wang et al., 2009b) over the last 50 years has been reported for northern Eurasia (60–40°N) with a possible northward shift with increased cyclone frequency in the higher latitudes (50–45°N) and decrease in the lower latitudes (south of 45°N), based on a study with reanalyses. The low latitude (south of 45°N) decrease was also noted by Zou et al. (2006) who reported a decrease in the number of severe storms for mainland China based on an analysis of extremes of observed 6-hourly pressure tendencies over the last 50 years.

Alexander and Power (2009) showed that the number of observed severe storms at Cape Otway (south-east Australia) has decreased significantly since the mid-19th century, strengthening the evidence of a southward shift in Southern Hemisphere storm tracks previously noted using reanalyses (Fyfe, 2003; Hope et al., 2006). Frederiksen and Frederiksen (2007) linked the reduction in cyclogenesis at 30°S and southward shift to a decrease in the vertical mean meridional temperature gradient. Pezza et al. (2007) confirmed previous studies using reanalysis showing a trend towards fewer and more intense low pressure systems. A study (Lim and Simmonds, 2009) using the ERA-40 reanalysis instead of the NCEP reanalysis used in previous studies, confirmed the trend towards more intense systems, but did not support the decrease in the number of cyclones seen in previous studies, emphasising the weaker consistency among reanalysis products for the Southern Hemisphere extratropical cyclones. Recent studies support the notion of more cyclones around Antarctica when the Southern Annular Mode (SAM) is in its positive phase and a shift of cyclones toward midlatitudes when the SAM is in its negative phase (Pezza and Simmonds, 2008). Additionally, more intense (and fewer) cyclones seem to occur when the Pacific Decadal Oscillation (PDO) is strongly positive and vice versa (Pezza et al., 2007).

In conclusion, it is *likely* that there has been a poleward shift in the main northern and southern stormtracks, during the last 50 years. The degree of agreement is high between several reanalysis products and through a wide selection of cyclone parameters and cyclone identification methods and European and Australian pressure based storminess proxies are consistent with a poleward shift over the last 50 years indicating that the evidence is robust. Advances have been made in documenting the observed decadal and multidecadal variability of extratropical cyclones using proxies for storminess. So the recent poleward shift should be seen in light of new studies with longer time spans that indicate that the last 50 years coincide with relatively low cyclonic activity in northern coastal Europe in the beginning of the period. Several studies using reanalyses suggest an intensification of high latitude cyclones, but there is still insufficient knowledge of how changes in the observational systems are influencing the reanalyses so even when the level of agreement is high among the studies the evidence cannot be considered to be strong. Other regional changes in intensity and the number of cyclones have been reported. The level of agreement between different studies using different tracking algorithms, different reanalysis or different cyclone parameters is still low. Thus, our confidence in the amplitude, and in some regions the sign, of the regional changes is *low*.

Regarding possible causes of the observed poleward shift, the AR4 concluded that trends over recent decades in the Northern and Southern Annular Modes, which correspond to sea level pressure reductions over the poles, are *likely* related in part to human activity, but an anthropogenic influence on extratropical cyclones was not formally detected, owing to large internal variability and problems due to changes in observing systems (Hegerl et al., 2007).

Anthropogenic influence on the sea level pressure distribution has been detected in individual seasons (Giannini et al., 2003; Gillett et al., 2005; Wang et al., 2009d). The trend pattern in atmospheric storminess as inferred from geostrophic wind energy and ocean wave heights has been found to contain a detectable response to anthropogenic and natural forcings with the effect of external forcings being strongest in the winter hemisphere (Wang et al., 2009d). However, the models generally simulate smaller changes than observed and also appear to under-estimate the internal variability, reducing the robustness of their detection results. New idealized studies have advanced the physical understanding of how stormtracks may respond to changes in the underlying surface conditions, indicating that a uniform SST increase weakens (reduced cyclone intensity or number of cyclones) and shifts the stormtrack poleward and strengthened SST gradients near the subtropical jet may lead to a meridional shift in the stormtrack either towards the poles or the equator depending on the location of the SST gradient change (Brayshaw et al., 2008), but the average global cyclone activity is not expected to change much under moderate greenhouse gas forcing (O'Gorman and Schneider, 2008; Bengtsson et al., 2009).

In summary, anthropogenic influence on the observed poleward shift in extratropical cyclones activity is *about as likely as not*. It has not formally been detected. However indirect evidence such as anthropogenic influence on the sea level pressure distribution and trend patterns in atmospheric storminess inferred from geostrophic wind and ocean wave heights has been found. In addition, increasing physical understanding that SST changes and greenhouse gas forcing influences extratropical cyclone storm tracks, enhances our confidence in this assessment.

The AR4 reports that in a future warmer climate, a consistent projection from the majority of the coupled atmosphere-ocean GCMs is fewer mid-latitude storms averaged over each hemisphere (Meehl et al., 2007b), a poleward shift of storm tracks in both hemispheres (particularly evident in the Southern Hemisphere), with greater storm activity at higher latitudes (Meehl et al., 2007b).

A northern hemisphere poleward shift in the upper tropospheric stormtrack due to increased anthropogenic forcing is supported by post-AR4 studies (Lorenz and DeWeaver, 2007). It should be noted that other studies indicate that the poleward shift is less clear when models including a full stratosphere and ozone recovery are used (Son et al., 2008) and the strength of the poleward shift is often seen more clearly in upper-level quantities than in low-level transient parameters (Ulbrich et al., 2008). Post-AR4 single model studies support the projection of a reduction in extratropical cyclones averaged over the northern hemisphere during future warming (Finnis et al., 2007; Bengtsson et al., 2009; Orsolini and Sorteberg, 2009). However, neither the global changes in storm frequency or intensity are found to be statistically significant by Bengsston et al. (2009), although they were accompanied by significant increases in total and extreme precipitation.

Models tend to project a reduction of wintertime cyclone activity throughout the mid-latitude North Pacific and for some models a northeastern movement of the North Pacific storm track (Loeptien et al., 2008; Ulbrich et al., 2008; Favre and Gershunov, 2009). However, the exact geographical pattern of cyclone frequency anomalies exhibits large variations across models (Teng et al., 2008; Laine et al., 2009). Over the North Atlantic stormtrack many models project an eastward or southeastward extension of the stormtrack (Ulbrich et al., 2008; Laine et al., 2009) with some models projecting a reduction in cyclone frequency along the Canadian east coast (Bengtsson et al., 2006; Watterson, 2006; Pinto et al., 2007b; Teng et al., 2008; Long et al., 2009). Using bandpassed MSLP from 16 CMIP3 coupled GCMs, Ulbrich et al. (2008) showed regional increases of the storm track activity over the Eastern North Atlantic/Western European area and Donat et al. (2010a) projected an increase in wind storm days for central Europe between 19 and 33% by the end of the 21st century (the increase varies according to the definition of storminess and one model projects a decrease) using 7 different coupled models. New results on Southern Hemisphere cyclones confirm the previously projected poleward shift in stormtracks under increased greenhouse gases (Lim and Simmonds, 2009). That study projected a reduction of Southern Hemisphere extratropical cyclone frequency and intensity in midlatitudes but a slight increase at high latitude.

Detailed analyses of changes in physical mechanisms related to cyclone changes in coupled climate models are still few. O'Gorman (2011) showed that changes in mean available potential energy of the atmosphere can account for much of the varied response in storm-track intensity to global warming implying that changes in storm-track intensity are sensitive to competing effects of changes in temperature gradients and static stability in different atmospheric levels. Using two coupled climate models, Laine et al. (2009) indicate that the primary cause for synoptic activity changes at the western end of the northern hemisphere storm tracks is related to the baroclinic conversion processes linked to mean temperature gradient changes in localized regions of the western oceanic basins. They also found downstream changes in latent heat release during the developing and mature stages of the cyclone to be of importance and indicated that changes in diabatic process may be amplified by the upstream baroclinic changes (stronger (weaker) baroclinic activity in the west gives stronger (weaker) latent heat release downstream). Pinto et al. (2009) found that regional increases in track density and intensity of extreme cyclones close to the British isles using a single model was associated with an eastward shift of the jet stream into Europe, more frequent extreme values of baroclinicity, and stronger upper level divergence. The modelled reduction in Southern Hemisphere extratropical cyclone frequency and intensity in the midlatitudes has been attributed to the tropical upper tropospheric warming enhancing static stability and decreasing baroclinicity while an increased meridional temperature gradient in the high latitudes is suggested to be responsible for the increase of cyclone activity in this region (Lim and Simmonds, 2009). In addition to details in the modelled changes in local baroclinicity and diabatic changes, the geographical pattern of modelled response in cyclone activity has been reported to likely be influenced by the individual model's structure of intrinsic modes of variability (Branstator and Selten, 2009).

In summary it is *likely* that there has been a poleward shift in the main northern and southern stormtracks, during the last 50 years. The degree of agreement is high between different studies and data indicating that the evidence is robust. Anthropogenic influence on this observed poleward shift in extratropical cyclone activity is *about as likely as not*. It has not formally been detected. It is *about as likely as not* that an increased anthropogenic forcing will give a reduction in the number of mid-latitude cyclones averaged over each hemisphere. Confidence in a poleward shift of the upper tropospheric storm tracks due to future anthropogenic forcings is only *medium*. Regional changes may be substantial and IPCC AR4 simulations show some regions with medium degree of agreement. However, studies using different analysis techniques, different physical quantities, different thresholds and different atmospheric vertical levels to represent cyclone activity and storm tracks result in different projections of regional changes. This leads to *low confidence* in region-specific projections.

3.5. Observed and Projected Impacts on the Natural Physical Environment

3.5.1. Droughts

Drought is generally "a period of abnormally dry weather long enough to cause a serious hydrological imbalance" (see IPCC SREX glossary and Box 3.2). While lack of precipitation (i.e., meteorological drought, Box 3.2) is often the primary cause of drought, increased evapotranspiration induced by e.g., enhanced temperature or radiation (e.g., Dai et al., 2004; Easterling et al., 2007; Corti et al., 2009), as well as preconditioning (pre-event soil moisture, lake, snow and/or groundwater storage) can contribute to the emergence of soil moisture and hydrological drought (Box 3.2). As noted in the AR4 (Trenberth et al., 2007), there are few direct observations of drought-related variables, in particular of soil moisture, available for a global analysis (see also Section 3.2.1). Hence, proxies for drought are often used to infer changes in drought conditions. Box 3.2 provides a discussion of the issue of drought definition and a description of commonly used drought indices. In order to understand the impact of droughts (e.g., on crop yields, general ecosystem functioning, water resources and electricity production), the timing, the duration, the intensity and the spatial extent need to be characterized. Other weather elements may interact to increase the impact of droughts: enhanced air temperature leads to enhanced evaporative demand, as does enhanced wind speed or increased incoming radiation. Moreover, climate phenomena such as monsoons (Section 3.4.1) and ENSO (Section 3.4.2) affect changes in drought occurrence in some regions. Hence, drought is a complex phenomenon that is strongly affected by other extremes considered in this Chapter. In addition, via land-atmosphere interactions, drought also has the potential to impact other weather and climate elements such as temperature and precipitation and associated extremes (Koster et al., 2004b; Seneviratne et al., 2006a ; Hirschi et al., 2011; see also Section 3.1.4).

The AR4 reported that very dry areas (PDSI < -3) had more than doubled in extent since 1970 on the global scale (Trenberth et al., 2007). However from a paleoclimate perspective recent droughts are not unprecedented, with severe "mega droughts" reported in the paleoclimatic record for Europe, North America and Australia. Recent studies extend this observation to African and Indian droughts (Sinha et al., 2007; Shanahan et al., 2009): much more severe and longer droughts occurred in the past centuries with widespread ecological political and socioeconomic consequences. Overall, these studies confirm that in the last millennium several extreme droughts (often associated with very warm air temperatures) have occurred (Breda and Badeau, 2008; Kallis, 2008; Büntgen et al., 2010); hence the current situation is not unprecedented.

START BOX 3.2 HERE

Box 3.2: The Definition of Drought

What is Drought or Dryness?
The IPCC SREX glossary defines drought as follows:
"A period of abnormally dry weather long enough to cause a serious hydrological imbalance. Drought is a relative term, therefore any discussion in terms of precipitation deficit must refer to the particular precipitation-related activity that is under discussion. For example, shortage of precipitation during the growing season impinges on crop production or

ecosystem function in general (due to soil moisture drought also termed agricultural drought), and during the runoff and percolation season primarily affects water supplies (hydrological drought). Storage changes in soil moisture and groundwater can also be strongly affected by evapotranspiration excesses in addition to precipitation deficits. A period with an abnormal precipitation deficit is defined as a meteorological drought. A megadrought is a very lengthy and pervasive drought, lasting much longer than normal, usually a decade or more".

As highlighted in the above definition, drought can be defined from different perspectives, depending on the involved stakeholders. The scientific literature commonly distinguishes *meteorological drought*, which refers to deficit of precipitation, *soil moisture drought* (often called *agricultural drought*), which refers to deficit of (mostly root zone) soil moisture, and *hydrological drought*, which refers to negative anomalies in streamflow, lake and/or groundwater levels (e.g., Heim Jr, 2002). We use here the term soil moisture drought instead of agricultural drought, despite the widespread use of the latter (e.g., Heim Jr, 2002; Wang, 2005), because soil moisture deficits have several additional effects beside those on agroecosystems, most importantly on other natural or managed ecosystems (including both forests and pastures), on building infrastructure (Corti et al., 2009), and health through impacts on heatwaves (Section 3.1.4).

Water scarcity which is caused additionally by overuse from human activities does not lie within the scope of this chapter (see Chapter 4); however it should be noted that increasing pressure on water resources by human uses may exert a positive feedback on drought e.g., via declining groundwater levels as a result of an intensive use of superficial and groundwater for agriculture. Drought should not be confounded with aridity, which describes the general characteristic of an arid climate. Indeed, drought is considered a recurring feature of climate occurring in any region and defined with respect to the average climate of the given region (e.g., Dai, 2011). Nonetheless, effects of droughts are not linear, given the existence of e.g., discrete soil moisture thresholds affecting vegetation and surface fluxes (e.g., Koster et al., 2004b; Seneviratne et al., 2010), which means that the same precipitation deficit or radiation excess relative to normal will not affect different regions equally. In this chapter we often use the term "*dryness*" instead of "drought" as a more general qualifier.

Drought Drivers

From a surface perspective (soil moisture or hydrological droughts), the two main drivers for droughts are precipitation deficit and/or evapotranspiration excess (Box 3.2, Figure 1). There are few examples of systems uniquely affected by precipitation deficits. Because soil moisture, groundwater and surface waters are associated with water storage, they have a characteristic memory (e.g., Vinnikov et al., 1996; Eltahir and Yeh, 1999; Koster and Suarez, 2001; Seneviratne et al., 2006b) and thus specific response times to drought forcing (e.g., Begueria et al., 2010). Furthermore, the memory is also a function of the atmospheric forcing and system's feedbacks (Koster and Suarez, 2001; Wang et al., 2009a), and the relevant storage is dependent on soil characteristics and rooting depth of the considered ecosystems. This means that drought has a different persistence depending on the affected system, and that it is also sensitive to pre-conditioning (Box 3.2, Figure 1). Effects of pre-conditioning also explain the possible occurrence of multi-year droughts, whereby soil moisture anomalies can be carried over from one year to the next (e.g., Wang, 2005). However, other features can induce drought persistence, such as persistent circulation anomalies, possibly strengthened by land-atmosphere feedbacks (Schubert et al., 2004). The choice of variable (e.g., precipitation, soil moisture, or streamflow) and time scale can strongly affect the ranking of drought events (Vidal et al., 2010).

[INSERT BOX 3.2, FIGURE 1 HERE
Box 3.2, Figure 1: Processes and drivers relevant for meteorological, soil moisture, and hydrological droughts.]

Drought Indicators

Because of the complex definition of droughts, and the lack of soil moisture observations (Section 3.2.1), several proxies have been developed to characterize (meteorological, soil moisture, and hydrological) drought (see e.g., Heim Jr, 2002; Dai, 2011). These proxies include (land-surface, hydrological or climate) model simulations (providing estimates of e.g., soil moisture or runoff), indices based on measured meteorological or hydrological variables, and paleoclimate proxies such as tree rings, speleothems or historical evidence such as harvest dates. We provide here a brief overview on the wide range of drought indices used in the literature.

Some indices are purely based on precipitation data. A widely used index is the Standard Precipitation Index (SPI) (McKee et al., 1993; Lloyd-Hughes and Saunders, 2002), which consists of fitting and transforming a long-term precipitation record into a normal distribution that has zero mean and unit standard deviation. SPI values of -0.5 to -1 correspond to mild droughts, -1 to -1.5 to moderate droughts, -1.5 to -2 to severe droughts and below -2 to extreme droughts. Similarly, values from 0 to 2 correspond to mildly wet to severely wet conditions, and values above 2 to extremely wet conditions (Lloyd-Hughes and Saunders, 2002). The SPI can be computed over several time scales (e.g., 3, 6 or 12 months) and thus indirectly considers effects of accumulating precipitation deficits, which are critical for soil moisture and hydrological droughts. Another index commonly used in the analysis of climate model simulations is the Consecutive Dry Days (CDD) index, which considers the maximum consecutive number of days without rain (i.e., below a given threshold, typically 1mm/day) within a considered period (i.e., year in general; Frich et al., 2002;

Alexander et al., 2006; Tebaldi et al., 2006). Though the SPI and CDD are both only based on precipitation, they do not necessarily only consider the effects of meteorological drought, since periods without rain are bound to have higher radiation forcing and thus possibly positive evapotranspiration anomalies. This is particularly the case because they focus on indirect precipitation characteristics (normalized precipitation anomalies for the SPI, duration of rainfree period for CDD), which do not relate the drought index to a specific precipitation amount.

Some indices consider both precipitation and evapotranspiration forcing, using simple parameterizations for the evapotranspiration estimates. These include the Palmer Drought Severity Index (PDSI) (Palmer, 1965), which measures the departure of moisture balance from normal conditions using a simple water balance model (e.g., Dai, 2011), as well as other indices such as the Precipitation Potential Evaporation Anomaly (PPEA) used in Burke and Brown (2008) and the Standardised Precipitation-Evapotranspiration Index (SPEI) described in Vicente-Serrano et al. (2010). The PDSI has been widely used for decades (in particular in the United States), also in climate-change analyses (e.g., Burke and Brown, 2008; Dai, 2011). The PDSI was originally calibrated for the central United States, which can impair the comparability of the index across regions, and thus it is often of advantage to renormalize the local PDSI (Dai, 2011). This can also be done using the self-calibrated PDSI (Wells et al., 2004). The model underlying the PDSI is essentially a simple bucket model, which is less advanced than more recent land surface and hydrological models.

For the assessment of soil moisture drought, also simulated soil moisture can be considered. Although the soil moisture simulated by (land-surface, hydrological and climate) models often exhibit strong discrepancies in absolute terms, soil moisture anomalies can be compared with simple scalings and generally match reasonably well (e.g., Koster et al., 2009; Wang et al., 2009a). Climate-change studies considering modeled soil moisture include those by Wang (2005), Sheffield and Wood (2008a), Wang et al. (2009a), and Orlowsky and Seneviratne (2011).

Other drought indices are used to quantify hydrological drought (e.g., Heim Jr, 2002; Vidal et al., 2010; Dai, 2011), but are less commonly used in the context of climate-change studies. Further analyses or indices also consider the area affected by droughts (e.g., Sheffield and Wood, 2008a; Dai, 2011) or additional variables (such as snow or vegetation indices from satellite measurements, e.g., Heim Jr, 2002).

Summary

In summary, drought indices often integrate temperature, precipitation and other variables, but may emphasize different aspects of drought and should be carefully selected with respect to the drought characteristic in mind (e.g., Nicholls and Alexander, 2007). For this reason, assessments of changes in drought characteristics with climate change should consider several indices to allow robust conclusions.

END BOX 3.2 HERE

There are still large uncertainties regarding observed global-scale trends in droughts. Globally, increases in the land area affected by drought was identified in two studies based on the PDSI model (Dai et al., 2004; Burke et al., 2006). This trend in the PDSI proxy was found to be largely affected by changes in temperature, not precipitation. On the other hand, based on soil moisture simulations with an observation-driven land surface model for the time period 1950-2000, Sheffield and Wood (2008a) have inferred trends in drought duration, intensity and severity predominantly decreasing, but with strong regional variation (and including increases in some regions). They concluded that there was an overall wetting trends over the considered time period, but also a switch since the 1970s to a drying trend, globally and in many regions, especially in high northern latitudes. Some regional studies are consistent with the results from Sheffield and Wood (2008a), regarding e.g., less widespread increase (or statistically insignificant changes or decreases) in some regions compared to the study of e.g., Dai et al. (2004) (e.g., in Europe, see below). More recently, Dai (2011) by extending the record did, however, find widespread increases in drought both based on various versions of the PDSI (for 1950-2008) and soil moisture output from a land surface model (for 1948-2004). Hence there are still large uncertainties with respect to global assessments of past changes in droughts. Nonetheless, there is some agreement between studies regarding increasing drought occurrence in some regions, although other regions also indicate opposite trends. Table 3.2 provides regional and continental-scale assessments of observed trends in dryness based on different indices (Box 3.2). The following paragraphs provide more details by continents.

In North America, there is *medium confidence* that there has been an overall slight tendency towards less dryness (wetting trend with more soil moisture and runoff; Table 3.2), although analyses for some subregions also indicate tendencies towards increasing dryness. This assessment is based on several lines of evidence, including simulations with different hydrological models as well as PDSI and CDD estimates (Alexander et al., 2006; Andreadis and Lettenmaier, 2006; van der Schrier et al., 2006a; Kunkel et al., 2008; Sheffield and Wood, 2008a; Wang et al., 2009c; Dai, 2011). The most severe droughts in the 20th century have occurred in the 1930s and 1950s, where the 1930s Dust Bowl was most intense and the 1950s drought most persistent (Andreadis et al., 2005) in the U.S., while in Mexico the 1950s and late 1990s were the driest periods. Recent regional trends towards more severe drought conditions were identified over southern and western Canada, Alaska and Mexico.

In Europe, there is *medium confidence* regarding increases in dryness based on some indices in the southern part of the continent, but large inconsistencies between indices in this region, and inconsistent or statistically insignificant trends in the rest of the continent (Table 3.2). Although Dai et al. (2004) found an increase in dryness for most of the European continent based on the PDSI, Lloyd-Hughes and Saunders (2002) and van der Schrier et al. (2006b) concluded, based on the analysis of SPI and self-calibrating PDSI for the 20th century (for 1901-1999, and 1901-2002, respectively), that no statistically significant changes were observed in extreme and moderate drought conditions in Europe (with the exception of the Mediterranean in van der Schrier et al., 2006b). Sheffield and Wood (2008a) also found contrasting dryness trends in Europe, with increases in the southern and eastern part of the continent, but decreases elsewhere. Beniston (2009b) reported a strong increase in warm-dry conditions over all central-southern (incl. maritime) Europe via a quartile-analysis from mid- to the end of the 20th century. Alexander et al. (2006) found trends towards increasing CDD mostly in the southern and central part of the continent, Trends of decreasing precipitation and discharge are consistent with increasing salinity in the Mediterranean, indicating a trend towards fresh water deficits (Mariotti et al., 2008), but this could also be partly caused by increased human water use. In France, an analysis based on a variation of the PDSI model also reported a significant increasing trend in drought conditions, in particular from the 1990s onward (Corti et al., 2009). Stahl et al. (2010) investigated streamflow data across Europe and found negative trends (lower streamflow) in southern and eastern regions, and generally positive trends (higher streamflow) elsewhere (especially in northern latitudes). Low flows have decreased in most regions where the lowest mean monthly flow occurs in summer, but vary for catchments which have flow minima in winter and secondary low flows in summer. The exceptional 2003 summer heat wave on the European continent (see Section 3.3.1) was also associated with a major soil moisture drought, as could be inferred from satellite measurements (Andersen et al., 2005), model simulations (Fischer et al., 2007a; 2007b), and impacts on ecosystems (Ciais et al., 2005; Reichstein et al., 2007).

There is *low to medium confidence* in dryness trends in South America (Table 3.2). For the Amazon, repeated intense droughts have been occurring in the last decades but no particular trend has been reported. The 2005 drought in Amazonia is, however, considered the strongest in the last century both from precipitation records and water storage estimates via satellite (measurements from the Gravity Recovery and Climate Experiment, (Chen et al., 2009)). For other parts of South America analyses of the return intervals between droughts in the instrumental and reconstructed precipitation series indicate that the probability of drought has increased during the late 19th and 20th centuries, consistent with selected long instrumental precipitation records and with a recession of glaciers in the Chilean and Argentinian Andean Cordillera (Le Quesne et al., 2006; 2009).

Changes in drought patterns have been reported for the monsoon regions of Asia and Africa with variations at the decadal timescale (e.g., Janicot, 2009). In Asia there is overall *low confidence* in trends in dryness both at the continental and regional scale, mostly due to spatially varying trends, except in East Asia where a range of studies, based on different indices, show increasing dryness in the second half of the 20th century (Table 3.2). In the Sahel, recent years are characterized by a greater interannual variability than the previous 40 years (Ali and Lebel, 2009; Greene et al., 2009), and by a contrast between the western Sahel remaining dry and the eastern Sahel returning to wetter conditions (Ali and Lebel, 2009). Giannini et al. (2008) report a drying of the African monsoon regions, related to warming of the tropical oceans, and variability related to the El Niño–Southern Oscillation. There is overall *low* to *medium confidence* regarding regional dryness trends in Africa (Table 3.2).

For Australia Sheffield and Wood (2008a) only found very limited increases in dryness from 1950-2000 based on soil moisture simulated using existing climate forcing (mostly in southeastern Australia) and some marked decreases in dryness in Central Australia and the northwestern part of the continent. Dai (2011), for an extended period until 2008 and using different PDSI variants as well as soil moisture output from a land surface model, found a more extended drying trend in the eastern half of the continent, but also a decrease in dryness in most of the western half. Jung et al. (2010) inferred from a combination of remote sensing and global eddy covariance flux observations that in particular the decade after 1998 became drier in Australia (and parts of Africa and South America), leading to decreased evapotranspiration, but it is not clear if this is a trend or just decadal variation.

In conclusion, following the assessment of the AR4 which was largely based on one study, subsequent work has drawn a more differentiated picture both regionally and temporally. There is not enough evidence at present to suggest high confidence in observed trends in dryness due to lack of direct observations, some geographical inconsistencies in the trends, and some dependencies of inferred trends on the index choice. There is *medium confidence* that since the 1950s some regions of the world have experienced more intense and longer droughts (e.g., southern Europe, West Africa, East Asia) but also opposite trends exist in other regions (e.g., Central North America, Northwestern Australia).

The AR4 (Hegerl et al., 2007) concluded that it is *more likely than not* that anthropogenic influence has contributed to the increase in the droughts observed in the second half of the 20th century. This assessment was based on multiple lines of evidence including a detection study which identified an anthropogenic fingerprint in a global PDSI data set with high significance (Burke et al., 2006).

There is now a better understanding of the potential role of land-atmosphere feedbacks versus SST forcing for droughts (e.g., Schubert et al., 2008a; 2008b) as well as of potential impacts of land use changes (Deo et al., 2009), but large uncertainties remain in the field of land surface modelling and land-atmosphere interactions, in part due to lack of observations (Seneviratne et al., 2010) and inter-model discrepancies (Koster et al., 2004b; Dirmeyer et al., 2006; Pitman et al., 2009). Nonetheless, a new set of climate modelling studies show that U.S. drought response to SST variability is consistent with observations (Schubert et al., 2009). Inferred trends in drought are also consistent with trends in global precipitation and temperature, and the latter two are consistent with expected responses to anthropogenic forcing (Hegerl et al., 2007; Zhang et al., 2007a). The change in the pattern of global precipitation in the observations and in model simulations are also consistent with theoretical understanding of hydrological response to global warming that wet regions become wetter and dry regions drier in a warming world (Held and Soden, 2006). However, the 2005/2006 U.S. drought in the southeastern U.S. was different from what would be expected from model projected anthropogenic climate change in this region: The drought was caused by a reduction in precipitation (with simultaneous reduction in evaporation), but models project an increase in precipitation minus evaporation (Seager et al., 2009). For soil moisture and streamflow drought it has been suggested that the stomatal “antitranspirant” responses of plants to rising atmospheric CO_2 may lead to a decrease in evapotranspiration (Gedney et al., 2006). This could mean that increasing CO_2 levels alleviate soil moisture and streamflow drought, but this result is still debated. Hence, though these new studies have improved the understanding of the mechanisms leading to drought, there is still not enough evidence to alter the AR4 assessment, in particular given the associated observational data issues (Section 3.2.1). We thus assess that there is *medium confidence* (see also Section 3.1.5) that anthropogenic influence has contributed to the increase in the droughts observed in the second half of the 20th century.

The AR4 assessed that projections at the time indicated an increase in droughts in particular in subtropical and mid-latitude areas (Christensen et al., 2007). An increase in dry spell length and frequency was considered *very likely* over the Mediterranean area, southern areas of Australia and New Zealand and *likely* over most subtropical regions, with little change over northern Europe. Continental drying and the associated risk of drought was considered *likely* to increase in summer over many mid-latitude continental interiors (e.g., central and southern Europe, the Mediterranean), in boreal spring and dry periods of the annual cycle over Central America.

More recent global and regional climate simulations and hydrological models mostly support the projections from AR4, as summarized in the following paragraphs (see also Table 3.3), although we assess the overall confidence in drought projections as *medium* given the definitional issues associated with dryness (Box 3.2) and the partial lack of agreement in model projections when based on different dryness indices (see below). Indeed, particular care is needed in intercomparing ‘drought’ projections since very many different definitions are employed (corresponding to different types of droughts), from simple climatic indices such as CDD to more complex indices of soil moisture and hydrological drought (Box 3.2). A distinction also needs to be made between short-term and longer-term events. Blenkinsop and Fowler (2007), for example, demonstrate that while an RCM ensemble indicate an increase in short-term summer drought over most of the UK, the longer (multi-season) droughts are projected to become shorter and less severe (although uncertainties in the latter projections are large – see below). These various distinctions are generally not considered and most currently available studies only assess changes in very few (most commonly one or two) dryness indices.

On the global scale, Burke and Brown (2008) provided an analysis of projected changes in drought based on four indices (SPI, PDSI, PPEA and simulated soil moisture anomaly; for definitions see Box 3.2) using two model ensembles: one based on a GCM expressing uncertainty in parameter space, and a multi-model ensemble of 11 GCM simulations from the CMIP3. Their analysis revealed that SPI, based solely on precipitation, showed little change in the proportion of the land surface in drought, and that all the other indices, which include a measure of the atmospheric demand for moisture, showed a statistically significant increase with an additional 5%–45% of the land surface in drought. This study also highlighted large uncertainties in regional changes in drought. This is also consistent with the more recent analysis from Orlowsky and Seneviratne (2011) for projections of changes in two drought indices (CDD and simulated soil moisture) on the annual and seasonal (DJF and JJA) time scales based on a larger ensemble of (23) GCM simulations from the CMIP3 (Figure 3.10; 2080-2100 vs 1980-2000, A2 scenario). It can be seen that the two indices partly agree on some areas of increased drought (e.g., on the annual time scale, in the Mediterranean, Central Europe, Central North America, Southern Mexico, and South Africa). But some regions where the models show consistent increases in CDD (e.g., Australia, Northern Brazil) do not show consistent decreases in soil moisture. Conversely, regions displaying a consistent decrease of CDD (e.g., in Northeastern Asia) do not show a consistent increase in soil moisture. The large uncertainty of drought projections is particularly clear from the soil moisture projections, with e.g., no agreement among the models regarding the sign of changes in DJF in most of the globe. These results regarding changes in CDD and soil moisture are consistent with other published studies (Wang, 2005; Tebaldi et al., 2006; Burke and Brown, 2008; Sheffield and Wood, 2008b; Sillmann and Roeckner, 2008) and the areas that display consistent increasing drought tendencies for both indices have also been reported to display such tendencies for additional indices (e.g., Burke and Brown, 2008; Dai, 2011). Sheffield and Wood (2008b, their Figure 10) examined projections in drought frequency (for droughts of duration of 4-6 month and longer than 12 months, estimated from soil moisture anomalies) based on simulations with 8 GCMs and the SRES scenarios A2, A1B, and B1. They concluded that drought was projected to increase in several regions under these three scenarios (mostly consistent with those displayed in Figure 3.10 for soil moisture changes), although the projections of drought intensification were stronger for the more extreme emissions scenarios (A2 and A1B) than for the more moderate scenario (B1). Regions showing statistically significant increases in drought frequency were found to be broadly similar for all three scenarios, despite the more moderate signal in the B1 scenario (their Figures 8 and 9). This study also highlighted the large uncertainty of scenarios for drought projections, as scenarios were found to span a large range of changes in drought frequency in most regions, from close to no change to two- to three-fold increases (their Figure 10).

Regional climate simulations over Europe also highlight the Mediterranean region as being affected by more severe droughts, consistent with available global projections (Table 3.3; see also Giorgi, 2006; Beniston et al., 2007; Mariotti et al., 2008; Planton et al., 2008). Mediterranean (summer) droughts are projected to start earlier in the year and last longer. Also, increased variability during the dry and warm season is projected (Giorgi, 2006). One GCM-based study projected one to three weeks of additional dry days for the Mediterranean by the end of the century (Giannakopoulos et al., 2009). For North America, intense and heavy episodic rainfall events with high runoff amounts are interspersed with longer relatively dry periods with increased evapotranspiration, particularly in the subtropics. There is a consensus of most climate-model projections of a reduction of cool season precipitation across the U.S. southwest and northwest Mexico (Christensen et al., 2007), with more frequent multi-year drought in the American southwest (Seager et al., 2007). Reduced cool season precipitation promotes drier summer conditions by reducing the amount of soil water available for evapotranspiration in summer. For Australia, Alexander and Arblaster (2009) project increases in consecutive dry days, although consensus between models is only found in the interior of the continent. African studies indicate the possibility of relatively small scale (500 km) heterogeneity of changes in precipitation and drought, based on climate model simulations (Funk et al., 2008; Shongwe et al., 2009).

Global and regional studies of hydrological drought (Hirabayashi et al., 2008b; Feyen and Dankers, 2009) project a higher likelihood of streamflow drought by the end of this century, with a substantial increase in the number of drought days (defined as streamflow below a specific threshold) during the last 30 years of the 21st century over North and South America, central and southern Africa, the Middle East, southern Asia from Indochina to southern China, and central and western Australia. Some regions, including Eastern Europe to central Eurasia, inland China, and northern North America, project increases in drought. In contrast, wide areas over eastern Russia project a decrease in drought days. At least in Europe, streamflow drought is primarily projected to occur in the frost-free season.

[INSERT FIGURE 3.10 HERE

Figure 3.10: Projected annual and seasonal changes of two dryness indices: Consecutive dry days (CDD, days with pr < 1mm) and average soil moisture (mrso); CMIP3 projections, 2080-2100 time frame minus 1980-2000 time frame (A2 relative to 20C3M simulations), annual (top), DJF (middle) and JJA (bottom). Shading is only applied for areas where at least 66% of the models agree in the sign of the change; stippling is applied for regions where at least 90% of all models agree in the sign of the change [from Orlowsky and Seneviratne, 2011, after Tebaldi et al., 2006].]

Increased confidence in modelling drought stems from consistency between models and satisfactory simulation of drought indices during the past century (Sheffield and Wood, 2008a; Sillmann and Roeckner, 2008). Inter-model agreement is stronger for long-term droughts and larger spatial scales (in some regions, see above discussion), while local to regional and short-term precipitation deficits are highly spatially variable and much less consistent between models (Blenkinsop and Fowler, 2007). Lack of complete knowledge of the physical causes of meteorological droughts, and of the links to the large-scale atmospheric and ocean circulation, are still a source of uncertainty in drought simulations and projections. For example, plausible explanations have been proposed for projections of both a worsening drought and a substantial increase in rainfall in the Sahara (Biasutti and Sobel, 2009). Another example is illustrated with the relationship of rainfall in southern Australia with sea surface temperatures (SSTs) around northern Australia. On annual time scales, low rainfall is associated with cooler than normal SSTs. Yet the warming observed in SST over the past few decades has not been associated with increased rainfall, but with a trend to more drought-like conditions (Nicholls, 2009).

There are still further sources of uncertainties affecting the projections of trends in meteorological drought for the coming century. The two most important may be uncertainties in the development of the ocean circulation and feedbacks between land surface and atmospheric processes. These latter processes are related to the effects of drought on vegetation physiology and dynamics (e.g., affecting canopy conductance, albedo and roughness), with resulting (positive or negative) feedbacks to precipitation formation (Findell and Eltahir, 2003a, b; Koster et al., 2004b; Cook et al., 2006; Hohenegger et al., 2009; Seneviratne et al., 2010; van den Hurk and van Meijgaard, 2010), and possibly - as only recently highlighted – also feedbacks between droughts, fires and aerosols (Bevan et al., 2009). Furthermore, the development of soil moisture that results from complex interactions of precipitation, water storage as soil moisture (and snow), and evapotranspiration by vegetation, is still associated with large uncertainties, in particular because of lack of observations of soil moisture and evapotranspiration (Section 3.2.1), and issues in the representation of soil moisture-evapotranspiration coupling in current climate models (Dirmeyer et al., 2006; Seneviratne et al., 2010). Uncertainties regarding soil moisture-climate interactions are also due to uncertainties regarding the behaviour plant transpiration, growth and water-use efficiency under enhanced atmospheric CO_2 concentrations, which could potentially have major impacts on the hydrological cycle (Betts et al., 2007), but are not well understood yet (Hungate et al., 2003; Piao et al., 2007; Bonan, 2008; Teuling et al., 2009). The space-time development of hydrological drought as a response to a meteorological drought and the associated soil moisture drought (drought propagation, e.g., Peters et al., 2003) also needs more attention. There is some understanding of these issues at the catchment scale (e.g., Tallaksen et al., 2009), but these need to be extended to the regional and continental scales. This would lead to better understanding of the spotted maps of hydrological droughts, which would contribute to a better identification and attribution of droughts and help to improve global hydrological models and land surface models.

In summary, there is *medium confidence* that since the 1950s some regions of the world have experienced more intense and longer droughts, in particular in southern Europe, West Africa, and East Asia, but also opposite trends exist in other regions (e.g., Central North America and Northwestern Australia). New studies have improved the understanding of the mechanisms leading to drought. Post-AR4 studies indicate that there is *medium confidence* in a projected increase of duration and intensity of soil moisture and hydrological drought in some regions of the world, in particular in the Mediterranean, Central North America, and Southern Africa. Definitional issues and lack of data preclude higher confidence than *medium i*n observations of drought changes, while these issues plus the inability of models to include all the factors likely to influence droughts preclude stronger confidence than *medium* in the projections.

3.5.2. Floods

A flood is “the overflowing of the normal confines of a stream or other body of water, or the accumulation of water over areas that are not normally submerged” (glossary of the American Meteorological Society). Floods include river floods, flash floods, urban floods, pluvial floods, sewer floods, coastal floods, and glacial lake outburst floods. The main causes of floods are intense and/or long-lasting precipitation, snow/ice melt, a combination of previous types, dam break (e.g., glacial lakes), reduced conveyance due to ice jams or landslides, or by a local intense storm (Smith and Ward, 1998). Floods are affected by various characteristics of precipitation, such as intensity, duration, amount, timing, phase (rain or snow). They are also affected by drainage basin conditions such as water levels in the rivers, presence of snow and ice, soil character and status (frozen or not, soil moisture content and vertical distribution), rate and timing of snow/ice melt, urbanisation, existence of dikes, dams, and reservoirs (Bates et al., 2008). Along coastal areas flooding may be associated with storm surge events (Section 3.5.5). A change in the climate physically changes many of the factors affecting floods (e.g., precipitation, snow cover, soil moisture content, sea level, glacial lake conditions) and thus may consequently change the characteristics of floods. Engineering developments such as dikes and reservoirs regulate flow, and land use may also affect floods. Therefore the assessment of causes of changes in floods is complex and difficult. This chapter focuses on the spatial, temporal and seasonal changes in high flows and peak discharge in rivers related to climate change. River discharge simulation under a changing climate scenario requires a set of GCM or RCM outputs (e.g., precipitation and surface air temperature) and a hydrological model. A hydrological model may consist of a land surface model of GCM or RCM and a river routing model. Different hydrological models may yield quantitatively different river discharge, but in general they do not yield different signs of the trend if the same GCM/RCM outputs are used. So the ability of models to simulate floods, in particular regarding the signs of the past and future trends, largely depends on the ability of GCM/RCM to simulate precipitation changes. The ability of GCM/RCM to simulate temperature is important for river discharge simulation in snowmelt- and glacier-fed rivers. More details on the ability and uncertainties in hydrological projections are described later in this subsection. The impact of floods on human society and ecosystems and related changes are discussed in Chapter 4. Coastal floods are described as a part of the section on extreme sea level (Section 3.5.3) and coastal impacts (Section 3.5.5). Glacial lake outburst floods are discussed in Section 3.5.6. Literature on the impact of climate change on pluvial floods is scarce, but the changes in heavy precipitation discussed in Section 3.3.2 may imply changes in pluvial floods in some regions.

Worldwide instrumental records of floods at gauge stations are limited in spatial coverage and in time, and only a limited number of gauge stations spans more than 50 years, and even fewer more than 100 years (Rodier and Roche, 1984). However, this can be overcome partly by using pre-instrumental flood data from documentary records (archival reports, in Europe continuous over the last 500 yrs) (Brazdil et al., 2005), and from geological indicators of paleofloods (sedimentary and biological records over centuries to millennia scales) (Kochel and Baker, 1982). Analysis of these pre-instrumental flood records have revealed that (1) flood magnitude and frequency are very sensitive to subtle alterations in atmospheric circulation, with greater sensitivity for the largest “rare” floods (50-year flood and higher) than for smaller frequent floods (2-year floods) (Knox, 2000; Redmond et al., 2002); (2) high interannual and interdecadal variability is found in flood occurrences both in terms of frequency and magnitude although in most cases, cyclic or clusters of flood occurrence are observed in instrumental (Robson et al., 1998), historical (Vallve and Martin-Vide, 1998; Benito et al., 2003; Llasat et al., 2005) and paleoflood records (Ely et al., 1993; Benito et al., 2008); (3) past flood records may contain analogues of unusual large floods, similar to some recorded recently, sometimes claimed to be the largest on record. For example, pre-instrumental flood data shows that the 2002 summer flood in the Elbe did not reach the highest flood levels recorded in 1118 and 1845 although it was higher than other disastrous floods of 1432, 1805, etc. (Brázdil et al., 2006). However, the currently available pre-instrumental flood data is also limited.

The AR4 concluded that no gauge-based evidence had been found for a climate-related trend in the magnitude/frequency of floods during the last decades (Rosenzweig et al., 2007). However, it also pointed to possible changes that may imply trends in flood occurrence with climate change. For instance, Trenberth et al. (2007) highlighted a catastrophic flood that occurred along several central European rivers in 2002, although no flood nor mean precipitation trends could be identified in this region; however there was a trend to increasing precipitation variability which itself could imply an enhanced probability of flood occurrence. Regarding the spring flow peak, the AR4 concluded with high confidence that abundant evidence was found for an earlier occurrence in snowmelt- and glacier-fed rivers (Rosenzweig et al., 2007; Bates et al., 2008), though we expressly note here that a change in flow peak does not necessarily imply nor preclude changes in flood magnitude or frequency in the affected regions.

Although trends in flood magnitude/frequency might be expected in regions where temperature change affects snowmelt or ice cover (in particular northern high-latitude and polar regions), widespread evidence of such changes is not available. For example, there is no evidence of widespread common trends in the magnitude of extreme floods based on the daily river discharge of 139 Russian gauge stations for the last few to several decades, though a significant shift of spring discharge to earlier dates has been found (Shiklomanov et al., 2007). Lindström and Bergström (2004) mentioned that it is difficult to conclude that flood levels are increasing in the analysis of runoff trends in Sweden for 1807-2002.

In the U.S. and Canada during the 20th century and in the early 21st century, there is no compelling evidence for changes in the magnitude/frequency of floods (Lins and Slack, 1999; Douglas et al., 2000; McCabe and Wolock, 2002; Cunderlik and Ouarda, 2009; Villarini et al., 2009). There are relatively abundant studies on the changes and trends for rivers in Europe such as rivers in Germany and its neighbouring regions (Mudelsee et al., 2003; Tu et al., 2005; Yiou et al., 2006; Petrow and Merz, 2009), in the Swiss Alps (Allamano et al., 2009), in France (Renard et al., 2008), in Spain (Benito et al., 2005), and in the UK (Robson et al., 1998; Hannaford and Marsh, 2008), but a continental-scale assessment of the changes in the flood magnitude/frequency for Europe is difficult to provide because geographically organized patterns are not seen in the reported changes.

The number of analyses based on stream gauge records for rivers in other parts of the world is limited. Available (limited) analyses for Asia suggest the following changes: the annual flood maxima of the lower Yangtze region show an upward trend over the last 40 years (Jiang et al., 2008), the likelihood for extreme floods in the Mekong river has increased during the second half of the 20th century (Delgado et al., 2009), and both upward and downward trends are identified over the last four decades in four selected river basins of the northwestern Himalaya (Bhutiyani et al., 2008). In the Amazon region in South America, the 2009 flood set record highs in the 106 years of data for the Rio Negro at the Manaus gauge site in July 2009 (Marengo, 2011). However, such analyses cover only limited parts of the world. Evidence in the scientific literature from the other parts of the world, and for other river basins, appears to be very limited. For example, Conway et al. (2009) concluded that robust identification of hydrological change was severely limited by data limitations and other reasons for sub-Saharan Africa. Di Baldassarre et al. (2010) found no evidence that the magnitude of African floods has increased during the 20th century.

The above analysis indicates that research subsequent to the AR4 still does not show clear and widespread evidence of observed changes in the magnitude/frequency of floods at the global level based on instrumental records, and there is thus *low confidence* regarding the magnitude and even the sign of these trends. The main reason for this lack of confidence is due to *lack of literature and evidence*, since instrumental records of floods at gauge stations are limited in space and time, which limits the number of analyses. Pre-instrumental flood data can provide information for longer periods, but these data are even scarcer. There is abundant evidence for an earlier occurrence of spring peak river flows in snowmelt- and glacier-fed rivers (*high confidence*), though this feature may not necessarily be linked with (nor does preclude) changes in extreme floods in the concerned regions. Assessed observed changes in heavy precipitation events (Section 3.3.2) are, however, *likely* to imply past changes in (pluvial) flood occurrence in some regions.

The possible causes for changes in floods were assessed by Bates et al. (2008), a cross-cutting technical paper based on the AR4, but cause-and-effect between external forcing and changes in floods was not explicitly discussed nor in the AR4. More recent literature has, however, detected the influence of anthropogenically-induced climate change in variables that affect floods, such as aspects of the hydrological cycle including mean precipitation (Zhang et al., 2007a), heavy precipitation (see Section 3.3.2), and snowpack (Barnett et al., 2008), though a direct link to trends in floods is still not established. The influence of anthropogenically-induced climate change is nonetheless clearly detected in streamflow regimes in the western USA (Barnett et al., 2008; Hidalgo et al., 2009).

Many river systems are not in their natural state anymore, making it difficult to separate changes in the streamflow data that are caused by the changes in climate from those caused by human regulation of the river systems. River engineering and land use may have altered flood probability. Many dams are designed to reduce flood. Large dams have resulted in large scale land use change and may have changed the effective rainfall in some regions (Hossain et al., 2009).

In climates where seasonal snow storage and melting plays a significant role in annual runoff, the hydrologic regime is affected by changes in temperature. In a warmer world, a smaller portion of precipitation falls as snow (Hirabayashi et al., 2008a) and the melting of winter snow occurs earlier in spring, resulting in a shift in peak river runoff to winter and early spring. This has been observed in the western U.S. (Regonda et al., 2005; Clow, 2010) and in Canada (Zhang et al., 2001), along with an earlier breakup of river ice in Russian Arctic rivers (Smith, 2000). The observed trends toward earlier timing of snowmelt-driven streamflows in the western U.S. since 1950 are detectably different from natural variability (Barnett et al., 2008; Hidalgo et al., 2009). It is unclear if observed warming over several decades has affected the magnitude of the snowmelt flow peak, but projected warming may result either in an increase in spring flood peak (where winter snow depth increases, (Meehl et al., 2007b) or a decrease in spring flood peak (i.e., presumably because of decreased snow cover and amounts; (Hirabayashi et al., 2008b; Dankers and Feyen, 2009).

There is still a lack of studies identifying an influence of observed warming over several decades on rain-generated peak streamflow trends because of uncertainty in the observed streamflow data and low signal to noise ratio. However, limited evidence has emerged that anthropogenic warming may have influenced the likelihood of rainfall-dominated floods in some river basins in Europe (Pall et al., 2011). Overall, there is *low confidence* (*limited evidence*) that anthropogenic warming has affected the magnitude/frequency of floods, though it has detectably influenced several components of the hydrological cycle such as precipitation and snow melt, which may impact flood trends. The assessment of causes behind the changes in floods is inherently complex and difficult. Nevertheless, there is *high confidence* in that observed warming over several decades, that is attributable to anthropogenic forcing, has *likely* been linked to earlier spring flow peaks in snowmelt- and glacier-fed rivers (Rosenzweig et al., 2007; Bates et al., 2008), though this may not necessarily imply higher flood occurrence.

The number of studies that investigated projected flood changes in rivers especially at a regional or a continental scale was limited when the AR4 was published. A rare example introduced in the AR4 was the study by Milly et al.(2002) which based on monthly river discharge calculated from climate model outputs identified projected changes (mostly increases) in 'large' floods at selected extratropical river basins larger than 20,000 km^2. Projections of flood changes at the catchment/river-basin scale were also not abundantly cited in the AR4. Nevertheless, Bates et al. (2008) argued that projected increases in the frequency of heavy precipitation events (see also 3.3.2) would also imply an enhanced risk of rain-generated floods in the affected regions.

The number of regional- or continental-scale studies of projected changes in floods is still limited. Recently, a few studies for Europe (Lehner et al., 2006; Dankers and Feyen, 2008, 2009) and a study for the globe (Hirabayashi et al., 2008b) have indicated changes in the frequency and/or magnitude of floods in the 21st century at a large scale using daily river discharge calculated from RCM or GCM outputs and hydrological models. Most notable changes are projected to occur in northern and northeastern Europe in the late 21st century, but the results vary between studies. Three studies (Dankers and Feyen, 2008; Hirabayashi et al., 2008b; Dankers and Feyen, 2009) show a decrease in the probability of extreme floods, that generally corresponds to lower flood peaks, in northern and northeastern Europe because of a shorter snow season, while one study (Lehner et al., 2006) shows an increase in floods in the same region. For other parts of the world, Hirabayashi et al. (2008b) show an increase in the risk of floods in most humid Asian monsoon regions, tropical Africa and tropical South America.

Projections of flood changes at the catchment/river-basin scale are also not abundant in the scientific literature. Several studies have been undertaken for UK catchments (Cameron, 2006; Kay et al., 2009; Prudhomme and Davies, 2009) and catchments in continental Europe and North America (Graham et al., 2007; Thodsen, 2007; Leander et al., 2008; Raff et al., 2009; van Pelt et al., 2009). However, projections for catchments in other regions such as Asia (Asokan and Dutta, 2008; Dairaku et al., 2008), the Middle East (Fujihara et al., 2008), South America (Nakaegawa and Vergara, 2010), and Africa are rare. Most projections for rain-dominated catchments were carried out, and are being carried out, because climate models project rainfall intensification in regions where these catchments are located, which is anticipated to be a cause of more frequent or more severe floods. Flood probability is generally projected to increase in such catchments, but uncertainty is still large in the changes in the magnitude and frequency of floods (Cameron, 2006; Kay et al., 2009).

It has been recently recognized that the choice of GCMs is the largest source of uncertainties in hydrological projections if the same emission scenario is adopted, and uncertainties from downscaling methods are of secondary importance (Graham et al., 2007; Leander et al., 2008; Kay et al., 2009; Prudhomme and Davies, 2009), although, in general, hydrological-model projections require downscaling and bias-correction of GCM outputs (e.g., precipitation and temperature). The choice of hydrological models is also of secondary importance (Kay et al., 2009). Nevertheless, whether downscaling, bias-correction, and the choice of hydrological models are of secondary importance may depend on the selected region/catchment, the selected downscaling and bias-correction methods, and the selected hydrological models (Wilby et al., 2008). For example, the above mentioned inconsistency between the projections of flood changes in northern and northeastern Europe (Lehner et al., 2006; Dankers and Feyen, 2008; Hirabayashi et al., 2008b; Dankers and Feyen, 2009) has been considered to be primarily due to differences in the downscaling and bias-correction methods applied in the different studies (Dankers and Feyen, 2009). Downscaling and bias-correction are also a major source of uncertainty in rain-dominated catchments (van Pelt et al., 2009).

The number of projections of flood magnitude/frequency changes is still limited at regional and continental scales. Projections at the catchment/river-basin scale are also not abundant in the peer-reviewed scientific literature, especially for regions outside Europe and North America. In addition, considerable uncertainty remains in the projections of flood changes, especially regarding their magnitude and frequency. Therefore, our assessment is that there is *low confidence* (due to *limited evidence* as well as *to low agreement of projections)* in future projections of changes in flood magnitude and frequency derived from river discharge simulations. Nevertheless, an increase in the magnitude and/or frequency of rain-generated floods is anticipated in some catchments and regions where short-term (e.g., daily) rainfall extremes and/or long-term (e.g., monthly, wet-season total) rainfall extremes are projected to increase. This assessment is an extension of Bates et al. (2008) because we consider here several spatio-temporal scales in rain-generated floods. However we note that heavy precipitation as well as mean precipitation are projected to either decrease and/or increase depending on the regions considered (Section 3.3.2), and that changes in several variables (e.g., precipitation totals, frequency and intensity, snow cover and snow melt, soil moisture) are relevant for changes in floods. Confidence in change of one of these components alone may thus not be sufficient to confidently assess future changes in flood occurrence. The earlier shifts of spring peak flows in snowmelt- and glacier-fed rivers are robustly projected (Kundzewicz et al., 2007; Bates et al., 2008); thereby, the earlier shifts can be assessed as *very likely*, though this may not necessarily be relevant for flood occurrence. There is *low confidence* (*limited evidence and low agreement*) in the projected magnitude of the earlier peak flows in snowmelt- and glacier-fed rivers.

In summary, there is *low confidence* at the global level regarding observed changes in the magnitude and frequency of floods, and even the sign of such changes. There is *low confidence* in future projections of changes in flood magnitude and frequency. Nevertheless, an increase in the magnitude and/or frequency of rain-generated floods is anticipated in some catchments and regions where short-term (e.g., daily) rainfall extremes and/or long-term (e.g., monthly, wet-season total) rainfall extremes are projected to increase. Earlier spring peak flows in snowmelt and glacier-fed rivers are *very likely*, but there is low confidence in their projected magnitude.

3.5.3. Extreme Sea Levels

Extreme sea levels are caused by severe storms such as tropical or extratropical cyclones. The associated drop in atmospheric pressure and strong winds can produce storm surges at the coast, which may be further elevated by wave setup caused by an onshore flux of momentum due to wave breaking. Extreme sea levels can be expected to change in the future as a result of both mean sea level rise and changes in atmospheric storminess, neither of which will be spatially uniform across the globe. As discussed in sections 3.4.4 and 3.4.5, changes in the frequency or intensity of tropical and extratropical cyclones, and their location, may be expected, and these changes may differ between ocean basins. Variations in the rate of sea level rise will occur as a result of variations in heat content in the ocean which lead to different rates of thermal expansion (e.g., Bindoff et al., 2007; Church et al., 2010; Timmermann et al., 2010). In addition, rapid melting of ice sheets will lead to non-uniform rates of sea level rise across the globe due to adjustments in the Earth's gravitational field (e.g., Mitrovica et al., 2010).

Mean sea level has varied considerably over glacial time scales as the extent of ice caps and glaciers have fluctuated with global temperatures. Sea levels rose around 130 m since the last glacial maximum 20-25ka before present to around 7000 years ago and reached a level close to present at least 6000 years ago (Lambeck et al., 2010). As well as the influence on sea level extremes caused by rapidly changing coastal bathymetries (Clarke and Rendell, 2009) and large scale circulation patterns (Wanner et al., 2008), there is some evidence that changes in the behaviour of severe tropical cyclones has changed on centennial time scales which points to non-stationarity in extreme sea level events (Nott et al., 2009). Woodworth et al. (2011) use tide gauge records dating back to the 18th century, and saltmarsh data, to show that sea level rise has accelerated over this time frame.

The AR4 reported there was *high confidence* that the rate of observed sea level rise increased from the 19th to the 20th century (Bindoff et al., 2007). It also reported that the global mean sea level rose at an average rate of 0.17 [0.12 to 0.22] mm yr^{-1} over the 20th century, 1.8 [1.3 to 2.3] mm yr^{-1} over 1961 to 2003 and at a rate of 3.1 [2.4 to 3.8] mm yr^{-1} over 1993 to 2003. Whether the faster rate of increase during the latter period reflected decadal variability or an increase in the longer term trend was not clear. However there is increasing evidence that the contribution to sea level due to mass loss from Greenland and Antarctica is accelerating (Velicogna, 2009). The AR4 also reported that the rise in mean sea level and variations in regional climate led to a *likely* increase in trend of extreme high water worldwide in the late 20th century (Bindoff et al., 2007) and that it was *more likely than not* that humans contributed to the trend in extreme high sea levels (IPCC, 2007a). Since the AR4, Menendez and Woodworth (2010), using data from 258 tide gauges across the globe, has confirmed the earlier conclusions of Woodworth and Blackman (2004) that there was a trend in extreme sea levels globally, more pronounced since the 1970's, and this trend was consistent with trends in mean sea level (see also Lowe et al., 2010). A number of additional studies at particular locations support this finding (e.g., Marcos et al., 2009; Haigh et al., 2010).

Various studies also highlight the additional influence of climate variability on extreme sea level trends. Menendez and Woodworth (2010) report that the El Niño – Southern Oscillation (ENSO) has a large influence on interannual variations in extreme sea levels in the Pacific Ocean and the monsoon regions based on sea level records since the 1970s. In southern Europe, Marcos et al. (2009) report that changes in extremes are also significantly negatively correlated with the North Atlantic Oscillation (NAO). Ullmann et al. (2007) concluded that maximum annual sea levels in the Camargue had risen twice as fast as mean sea level during the 20th century due to an increase in southerly winds associated with a general rise in sea level pressure over central Europe (Ullmann et al., 2008). Sea level trends from two tide gauges on the north coast of British Columbia from 1939-2003 were twice that of mean sea level rise, the additional contribution being due to the strong positive phase of the Pacific Decadal Oscillation (PDO) which has lasted since the mid-1970s (Abeysirigunawardena and Walker, 2008). Cayan et al. (2008) reported increases in the frequency of exceedance of the 99.99th percentile sea level of 20-fold at San Francisco since 1915 and 30-fold at La Jolla since 1933, also noting that positive sea level anomalies of 10 to 20 cm that often persisted for several months during El Niño events produced an increase in storm surge peaks over this time. The spatial extent of these oscillations and their influence on extreme sea levels across the Pacific have been discussed by Merrifield et al. (2007). Church et al. (2006b) examined changes in extreme sea levels before and after 1950 in two tide gauge records of approximately 100 years on the east and west coasts of Australia respectively. At both locations a stronger positive trend was found in the sea level exceeded by 0.01 per cent of the observations than the median sea level, suggesting that in addition to mean sea level rise other modes of variability or climate change are contributing to the extremes. At Mar del Plata, Argentina, Fiore et al. (2009) noted an increase in the number and duration of positive storm surges in the decade 1996 to 2005 compared to previous decades which may be due to a combination of mean sea level rise and changes in wind climatology resulting from a southward shift in the South Atlantic high.

Studies since the AR4 conclude that trends in extreme sea level are generally consistent with changes in mean sea level (e.g., Marcos et al., 2009; Haigh et al., 2010; Menendez and Woodworth, 2010) although some studies note that the trends in extremes are larger than the observed trend in mean sea levels (e.g., Church et al., 2006b; Ullmann et al., 2007; Abeysirigunawardena and Walker, 2008) and may be influenced by modes of climate variability such as the PDO on the Canadian west coast (e.g., Abeysirigunawardena and Walker, 2008; Marcos et al., 2009; Menendez and Woodworth, 2010). These studies are consistent with the conclusions from the AR4 that increases in extremes are related to trends in mean sea level and modes of variability in the regional climate.

The AR4 (Meehl et al., 2007b) projected sea level rise for 2090–2099 relative to 1980–1999. The rise from ocean thermal expansion, glaciers and ice caps, and modelled ice sheet contributions is projected to be 18–59 cm which incorporates a 90% confidence range, across all scenarios. An additional allowance to the sea level rise projections was made for a possible rapid dynamic response of the Greenland and West Antarctic ice sheets, which could result in an accelerating contribution to sea level rise. This was estimated to be 10–20 cm of sea level rise by 2090-2099 using a simple linear relationship with projected temperature. Because of insufficient understanding of the dynamic response of ice sheets, Meehl et al. (2007b) also noted that a larger contribution could not be ruled out.

Several studies since the AR4 have developed statistical models that relate 20th century (e.g., Rahmstorf, 2007; Horton et al., 2008) or longer (e.g., Grinsted et al., 2009; Vermeer and Rahmstorf, 2009) temperature and sea level rise to extrapolate future global mean sea level. These alternative approaches yield projections of sea level rise by 2100 of 0.50-1.20 m (Rahmstorf, 2007), 0.47 - 1.00 m (Horton et al., 2008), 0.9 to 1.3 m for the A1B Scenario (Grinsted et al., 2009) and 0.75 – 1.90 m (Vermeer and Rahmstorf, 2009). However, as noted by Cazenave and Llovel (2010) future rates of sea level rise may be less closely associated with global mean temperature if ice sheet dynamics play a larger role in the future. Using glacier models, Pfeffer et al. (2008) found that sea level rise of more than 2 m by 2100 is physically implausible. An estimate of 0.8 m by 2100 that included increased ice dynamics was considered most plausible.

New studies, whose focus is on quantifying the effect of storminess changes on storm surge, have been carried out over the northern European region since the AR4 and mostly find an increase along the North Sea coastline. This is consistent with increased storminess and wind speed as indicated by most models across this region in Figure 3.9. Debernard and Roed (2008) investigated storm surge changes over Europe in four regionally downscaled GCMs including two run with B2, one with A2 and one with an A1B emission scenario. Despite large inter-model differences, statistically significant changes between 2071-2100 and 1961-1990 consisted of decreases in the 99th percentile surge heights south of Iceland, and an 8-10% increase along the coastlines of the eastern North Sea and the northwest British Isles, which occurred mainly in the winter season. Wang et al. (2008) projected a significant increase in wintertime storm surges around Ireland except the south Irish coast over 2031-2060 relative to 1961-1990 using a downscaled GCM under an A1B scenario. Sterl et al. (2009) concatenated the output from a 17 member ensemble of A1B simulations from a GCM over the model periods 1950-2000 and 2050-2100 into a single longer time series to estimate 10000 year return values of surge heights along the Dutch coastline. No statistically significant change in this value was projected for the 21st century because projected wind speed changes were not associated with the surge-generating northerlies but rather non-surge generating southwesterlies.

Other studies have undertaken a sensitivity approach to compare the relative impact on extreme sea levels of severe weather changes and mean sea level rise. Over southeastern Australia, McInnes et al. (2009b) found that a 10% increase in wind speeds, consistent with the upper end of the range under an A1FI scenario from a multi-model ensemble, would produce an increase in sea levels that were 20 to 35% of the upper end of the A1FI sea level rise projection for 2070. Brown et al. (2010) also investigated the relative impact of sea level rise and wind speed change on an extreme storm surge in the eastern Irish Sea. Both studies concluded that sea level rise rather than meteorological changes has the greater potential to increase extreme sea levels in the future.

The degree to which climate models (GCM or RCM) have sufficient resolution and/or internal physics to realistically capture the meteorological forcing responsible for storm surges is regionally dependant. For example current GCMs are unable to realistically represent tropical cyclones. This has led to the use of alternative approaches for investigating the impact of climate change on storm surges in tropical Australia whereby cyclone characteristics and tracks are represented by statistical models, from which populations of synthetic cyclones representing current climate can be constructed. Such models can also be perturbed to represent projected future climates (e.g., McInnes et al., 2003). Recent studies on the tropical east coast of Australia reported in Harper et al. (2009) that employ these approaches show a relatively small impact of a 10% increase in tropical cyclone intensity on the 1 in 100 year storm tide (the combined sea level due to the storm surge and tide), again with mean sea level rise producing the larger contribution to changes in future sea level extremes. However, one study that has incorporated scenarios of sea level rise in the hydrodynamic modelling of hurricane-induced sea level extremes on the Louisiana coast found that increased coastal water depths had a large impact on surge propagation, increasing storm surge heights by 2 to 3 times the sea level rise scenario, particularly in wetland-fronted areas (Smith et al., 2010).

To summarise, post-AR4 studies provide additional evidence that trends in extreme sea level across the globe reflect the trends in mean sea level, suggesting that mean sea level rise rather than changes in storminess are largely contributing to this increase (although data are sparse in many regions and this lowers the confidence in this assessment). It is considered *likely* that sea level rise has led to a change in extreme water levels. Studies into changes in future extreme sea levels have poor global coverage being mainly focussed on Europe although these studies generally provide further evidence for an increase in extreme sea levels due to changes in storminess in the North Sea. Other studies that have compared the relative contribution to future extreme sea levels of mean sea level and storminess changes find that, despite large uncertainties in the magnitude of these contributions in the future, sea level rise has been found to lead to larger increases in total water level, although the studies are limited in number and geographical coverage. On the basis of these studies of observed trends in extreme sea levels it is *very likely* that sea level rise will contribute to increases in extreme sea levels in the future. While changes in storminess may contribute to changes in sea level extremes, the limited geographical coverage of studies to date and the uncertainties associated with storminess changes overall (Sections 3.4.4 and 3.4.5) means that a general assessment of the effects of storminess changes on storm surge is not possible at this time.

3.5.4. Waves

Severe waves threaten the safety of coastal inhabitants and those involved in maritime activities and can damage and destroy coastal and marine infrastructure. Waves play a significant role in shaping a coastline by transporting energy from remote areas of the ocean to the coast. Energy dissipation via wave breaking contributes to beach erosion, longshore currents, and elevated coastal sea levels through wave set-up and wave run-up. Wave properties that influence these processes include wave height, the wave energy directional spectrum, and period, although to date studies of past and future wave climate changes have tended to focus on wave height parameters such as ‘significant wave height’ (SWH - the height from trough to crest of the highest one third of waves) and metrics of extreme waves, such as high percentiles or wave heights above particular thresholds. One study examines trends in SWH, mean wave direction and peak wave period (Dodet et al., 2010).

Wave climates have changed over paleo-climatic time scales. Wave modelling using paleobathymetries over the past 12000 years indicates an increase in peak annual SWH of around 40% due to the increase in relative sea level, which redefines the location of the coastline and hence progressively extends the fetch length in most of the shelf sea regions (Neill et al., 2009). Major circulation changes that result in changes in storminess and wind climate (see section 3.3.3) have also affected wave climates. Evidence of enhanced storminess determined from sand drift and dune building along the western European coast indicates that enhanced storminess occurred over the period of the little ice age (1570-1900) and the mid Holocene (~8200 BP; Clarke and Rendell, 2009).

The AR4 reported statistically significant positive trends in SWH over the period 1950 to 2002 over most of the mid-latitudinal North Atlantic and North Pacific, as well as in the western subtropical South Atlantic, the eastern equatorial Indian Ocean and the East China and South China Sea and declining trends around Australia, and parts of the Philippine, Coral and Tasman Seas (Trenberth et al., 2007), based on voluntary observing ship data (VOS; e.g., Gulev and Grigorieva, 2004). Several studies that address trends in extreme wave conditions have been completed since the AR4 and the new studies generally provide more evidence for the previously reported positive trends in SWH and extreme waves in the north Atlantic and north Pacific. Wang et al. (2009c) found that SWH increased in the boreal winter over the past half century in the high latitudes of the Northern Hemisphere (especially the northeast Atlantic), and decreased in more southerly northern latitudes based on ERA-40 reanalysis. They also found that storminess around the 1880s was of similar magnitude to that in the 1990s, consistent with Gulev (2004). In a regional North Sea hindcast, Weisse and Günther (2007) found a positive trend in 99th percentile wave height from 1958 to the early 1990s followed by a declining trend to 2002 over the southern North Sea, except on the UK North Sea coast where negative trends occurred over much of the hindcast period. A wave hindcast over the north-eastern Atlantic Ocean over the period 1953-2009 revealed a significant positive trend in SWH and a counterclockwise shift in mean direction in the north and a slight but not significant increase in peak wave period in the northeast. In the south, no trend was found for SWH or wave period while a clockwise trend in mean direction was found (Dodet et al., 2010). On the North American Atlantic coast, Komar and Allan (2008) found a statistically significant trend of 0.059 m/yr in waves exceeding 3 m during the summer months over 30 years since the mid-1970s at Charleston, South Carolina, with lower but statistically significant trends at wave buoys further north. These trends were associated with an increase in intensity and frequency of hurricanes over this period. In contrast, wintertime waves, generated by extratropical storms, were not found to have experienced a statistically significant change. Along the U.S. west coast, SWH is strongly correlated with El Niño. However positive trends were also found in SWH and extreme wave height from the mid-1970s to 2006 in wave buoy data (Allan and Komar, 2006), in hindcast SWH over 1948-1998 (Adams et al., 2008) and for excesses of the 98th percentile SWH over 1985-2007 (Menendez et al., 2008). Positive though not statistically significant trends in annual mean SWH were found over south-eastern South America for in situ wave data over the 1996-2006 period and in satellite wave data over 1993-2001 while simulated wave fields using reanalysis wind forcing over the period 1971-2005 produced statistically significant trends in SWH (Dragani et al., 2010). Trends at particular locations may be also influenced by local factors. For example, Suursaar and Kullas (2009) reported a slight decreasing trend in mean SWHs from 1966–2006 in the Gulf of Riga within the Baltic Sea, while the frequency and intensity of high wave events (i.e., the difference between the maximum and 99th percentile wave height) showed rising trends. These changes were associated with a decrease in local average wind speed, but an intensification of westerly winds and storm events occurring further to the west.

In the Southern Ocean SWH derived from satellite observations was found to be strongly positively correlated with SAM particularly from March to August (Hemer et al., 2010). However, the analysis of reliable long term trends in the Southern Hemisphere remains challenging due to limited in situ data and problems of temporal homogeneity in reanalysis products. For example, Hemer et al. (2010) also found that trends in SWH derived from satellite data over 1998–2000 relative to 1993–1996 were positive only over the Southern Ocean south of 45°S whereas trends were positive across most of the Southern Hemisphere in the corrected ERA-40 reanalysis (C-ERA-40). Furthermore, the frequency of wave events exceeding the 98th percentile over the period 1985-2002 using data from a wave buoy situated on the west coast of Tasmania showed no statistically significant trend whereas a strong positive trend was found in equivalent fields of C-ERA-40 data (Hemer, 2010).

New studies have demonstrated strong links between wave climate and natural modes of climate variability. For example, along the U.S. west coast and the western North Pacific, SWH was found to be strongly correlated with El Niño (Allan and Komar, 2006; Sasaki and Toshiyuki, 2007) and in the southern ocean, SWH was positivity correlated with the SAM. On the U.S. East coast, positive trends in summertime SWH were linked to increasing numbers of hurricanes. In the northeast Atlantic trends in SWH exhibited significant positive (negative) correlations with NAO in the north (south) and more generally, trends in SWH, mean wave direction and peak wave period over the period 1953-2009 were related to the increase in NAO index over this time (Dodet et al., 2010). One study (Wang et al., 2009d) reported a link between external forcing (i.e., anthropogenic forcing due to greenhouse gases and aerosols, and natural forcing due to solar and volcanic forcing) and an increase in SWH in the boreal winter in the high-latitudes of the Northern Hemisphere (especially the northeast North Atlantic), and a decrease in more southerly northern latitudes over the past half century.

The AR4 projected an increase in extreme wave height for many regions of the mid-latitude oceans due to a projected northward movement in storm tracks and associated increases in wind speeds in these regions due to projected increased greenhouse gas concentrations in the atmosphere (Meehl et al., 2007b). At the regional scale, increases in wave height were projected for most mid-latitude areas analysed, including the North Atlantic, North Pacific and Southern Ocean (Christensen et al., 2007) but with low confidence due to the low confidence in projected changes in mid-latitude storm tracks and intensities. Several studies since then have developed wave climate projections, which provide greater evidence for future wave climate change. Global scale projections of SWH were developed by Mori et al. (2010), using a 1.25° resolution wave model forced with projected winds from a 20 km global GCM, in which ensemble-averaged SST changes from the CMIP3 models provided the climate forcing. The spatial pattern of projected SWH change between 2075-2100 and 1979-2004 reflect the changes in the forcing winds, which are generally similar to the mean wind speed changes shown in Figure 3.9. Extreme waves (measured by the average of the top 10 values) were projected to exhibit large increases in the northern Pacific, particularly close to Japan due to an increase in strong tropical cyclones and also the Indian Ocean despite decreases in SWH.

A number of regional studies have also been completed since the AR4 in which forcing conditions were obtained for a few selected emission scenarios (typically B2 and A2, representing low-high ranges) from GCMs or RCMs. These studies provide additional evidence for positive trends in SWH and extreme waves along the western European coast (e.g., Debernard and Roed, 2008; Grabemann and Weisse, 2008), the UK coast (Leake et al., 2007), declines in extreme wave height in the Mediterranean (Lionello et al., 2008) and the southeast coast of Australia (Hemer et al., 2010) and little change along the Portuguese coast (Andrade et al., 2007). However, considerable variation in projections can arise from the different climate models and scenarios used to force wave models, which lowers the confidence in the projections. For example along the European North Sea coast, 99th percentile wave height over the late 21st century relative to the late 20th century is projected to increase by 6-8% by Debernard and Roed (2008) based on wave model simulations with forcing from several GCMs under A2, B2 and A1B greenhouse gas scenarios, whereas they are projected to increase by up to 18% by Grabemann and Weisse (2008) who downscaled two GCMs under A2 and B2 emission scenarios. In one region, opposite trends in extreme waves were projected. Grabemann and Weisse (2008) project negative trends in 99th percentile wave height along the UK North Sea coast, whereas Leake et al. (2007) downscaled the same GCM for the same emission scenarios, using a different RCM and found positive changes in high percentile wave heights offshore of the East Anglia coastline. Hemer et al. (2010) concluded that uncertainties arising from the method by which climate model winds were applied to wave model simulations (e.g., by applying bias-correction to winds or perturbing current climate winds with changes in winds derived from climate models) made a larger contribution to the spread of climate model projections than the forcing from different GCMs or emission scenarios.

In summary, although post-AR4 studies are few and their regional coverage is limited, their findings generally support the evidence from earlier studies of wave climate trends. Most studies find a link between variations in waves (both SWH and extremes) and internal climate variability. Only one study has detected a link between external forcing (anthropogenic and natural) and positive trends in SWH in northern high latitudes and negative trends in northern mid-latitudes. As a result, there is *low confidence* that there has been an anthropogenic influence on extreme wave heights. Additional downscaling studies in more climate model simulations provide further evidence for projected increases in wave height in some regions such as the eastern North Sea, but the small number of studies, the lack of consistency of the wind projections between GCMs combined with limitations in their ability to simulate extreme winds means there is *low confidence* in the findings. However the strong linkages between wave height and winds and storminess means that it is *likely* that future changes in SWH will reflect future changes in these parameters.

3.5.5. Coastal Impacts

Two classes of coastal hazard that are particularly significant in the context of disaster management are coastal inundation and shoreline stability both of which would be affected by climate change through rising sea levels and changes in extreme events. Coastal inundation occurs during periods of extreme sea levels due to storm surges and high waves, particularly when combined with high tides. Although tropical and extra-tropical cyclones are the most common causes of sea level extremes, other weather events that cause persistent winds such as anticyclones and fronts can also influence coastal sea levels (e.g., Green et al., 2009; McInnes et al., 2009b). In many parts of the world sea levels are influenced by modes of variability such as the El Niño – Southern Oscillation (ENSO, Section 3.4.2). In the western equatorial Pacific, sea levels can fluctuate up to half a metre between ENSO phases (Church et al., 2006a) and in combination with extremes of the tidal cycle, can cause extensive inundation in low-lying atoll nations in the absence of extreme weather events (Lowe et al., 2010). Shoreline position can change from the combined effects of various factors such as:

1. Rising mean sea levels, which causes landward recession of coastlines made up of erodible materials.
2. Subsidence of coastal terrain due to isostatic rebound (Blewitt et al., 2010; Mitrovica et al., 2010), or sediment compaction from the removal of oil, gas and water (Syvitski et al., 2009).
3. Changes in the frequency or severity of transient storm erosion events (Zhang et al., 2004a).
4. Changes in sediment supply to the coast (Stive et al., 2003; Nicholls et al., 2007; Tamura et al., 2010).
5. Changes in wave period due to sea level rise, which alters wave refraction, or in wave direction, which can cause cause realignment of shorelines (Ranasinghe et al., 2004; Bryan et al., 2008; Tamura et al., 2010).
6. The loss of natural protective structures such as coral reefs (e.g., Sheppard et al., 2005; Gravelle and Mimura, 2008) or the reduction of permafrost or sea ice in mid and high latitudes, which exposes soft shores to the effects of waves and severe storms (see 3.5.7, Manson and Solomon, 2007).

The susceptibility of coastal regions to erosion and inundation is related to various physical (e.g., shoreline slope), and geomorphologicical and ecosystem attributes, and therefore may be inferred to some extent from broad coastal characterisations. These include the presence of beaches, rocky shorelines or cliffed coasts; deltas; backbarrier environments such as estuaries and lagoons; the presence of mangroves, saltmarshes or sea grasses, shorelines flanked by coral reefs (e.g., Nicholls et al., 2007) or by permafrost or seasonal sea ice each of which are characterised by different vulnerability to climate change driven hazards. For example, deltas are low-lying and hence generally prone to inundation, while beaches are comprised of loose particles and therefore erodible. However, the degree to which these systems may be impacted by erosion and inundation may also be influenced by other factors which may affect disaster responses. For example, depleted mangrove forests or the degradation of coral reefs may lead to reduced protection from high waves during severe storms, (e.g., Gravelle and Mimura, 2008); there may be a loss of ecosystem services brought about by saltwater contamination of already limited freshwater reserves due to rising sea levels and these amplify the risks of climate change (McGranahan et al., 2007), and also reduce the resilience of coastal settlements to disasters. Dynamical processes such as vertical land movement also contributes to inundation potential (Haigh et al., 2009). Some coastal regions may be rising due to post-glacial rebound or slumping due to aquifer drawdown (Syvitski et al., 2009). Multiple contributions to coastal flooding such as heavy rainfall and flooding in coastal catchments that coincide with elevated sea levels may also be important. Ecosystems such as coral reefs also play an important role in providing material on which atolls are formed. Large scale oceanic changes that are particularly relevant to both coral reefs and small island countries are discussed in Box 3.3.

The rise in mean sea level by about 120 m since the end of the last ice age (Jansen et al., 2007) has had a profound effect on coastline position around the world. Contributing to the evolution of the coastlines have been changes in action of the ocean on the coast through changes in wave climate (Neill et al., 2009), tides (Gehrels et al., 1995) and changes in storminess (e.g., Clarke and Rendell, 2009).

The AR4 (Nicholls et al., 2007) reported that coasts are experiencing the adverse consequences of impacts such as increased coastal inundation, erosion and ecosystem losses. However, attributing these changes to sea level rise is difficult due to the multiple drivers of change over the 20th century (Nicholls, 2010) and the scarcity and fragmentary nature of data sets which contributes to the problem of identifying and attributing changes (e.g., Defeo et al., 2009). Since the AR4 there have been several new studies that examine coastline changes. In the Caribbean, the beach profiles at 200 sites across 113 beaches and eight islands were monitored on a three-monthly basis from 1985 to 2000 with most beaches found to be eroding and faster rates of erosion generally found on islands that had been impacted by a higher number of hurricanes. However, the relative importance of anthropogenic factors, climate variability and climate change on the eroding trends could not be separated quantitatively (Cambers, 2009). In Australia, Church et al. (2008) report that despite the positive trend in sea levels during the 20th century, beaches have generally been free of chronic coastal erosion, but where it has been observed, it has not been possible to unambiguously attribute it to sea level rise in the presence of other anthropogenic activities. A quantitative analysis of physical changes in 27 atoll islands across three central Pacific Nations (Tuvalu, Kiribati and Federated States of Micronesia) over a 19 to 61 year period using photography and satellite imagery found 43% of islands remained stable and 43% increased in area over the timeframe of analysis, with largest decadal rates of increase in island area ranging from 0.1 to 5.6 hectares. Only 14% of islands studied exhibited a net reduction in area (Webb and Kench, 2010). Despite the small net changes in area, a larger redistribution of land area, consisting of a net lagoonward migration of islands, was evident in 65% of cases. Chust et al. (2009) evaluated the relative contribution of local anthropogenic (non-climate change related) and sea level rise impacts on the coastal morphology and habitats in the Basque coast, northern Spain for the period 1954–2004. They found that the impact from local anthropogenic influences was about an order of magnitude greater than that due to sea level rise over this period.

START BOX 3.3 HERE

Box 3.3: Small Islands

Small islands represent a distinct category of locations owing to their small size and highly maritime climates, which means that their concerns and information needs in relation to future climate change differ in many ways from those of the larger continental regions that are addressed in this chapter. Particular challenges exist for the assessment of past changes of climate given the sparse regional and temporal coverage of terrestrial-based observation networks and the limited in situ ocean observing network although observations have improved somewhat in recent decades with the advent of satellite-based observations of meteorological and oceanic variables. However, the short length of these records hampers the investigation of long term trends in the region. The resolution of GCMs is insufficient to resolve small islands and few studies have been undertaken to provide projections for small islands using RCMs (Campbell et al., 2011). In regions such as the Pacific Ocean, large scale climate features such as the South Pacific Convergence Zone and the El Niño – Southern Oscillation (ENSO, Section 3.4.2) have significant influence on the pattern and timing of precipitation, yet these features and processes are often poorly represented in GCMs. The purpose of this box is to present available information on observed trends and climate change projections that are not covered in the other sections of this chapter as well as discuss key aspects of the climate system that are particularly relevant for small islands.

Although the underlying data sources are limited, some data for the Indian Ocean, South Pacific (Fiji) and Caribbean were available in the studies of Alexander et al. (2006) and Caesar et al. (2011). Problems of data availability and homogeneity for the Caribbean are discussed by Stephenson et al. (2008b). Based on standard extremes indices, positive trends in warm days and warm nights and negative trends in cool days and cool nights have occurred across the Indian Ocean and South Pacific region for the period 1971-2005 (Caesar et al., 2011) and the Caribbean for the period 1951-2003 (based on data from Alexander et al., 2006). Based on the same data sources, trends in average total wet day precipitation were positive and statistically significant over the Indian Ocean region, negative over the South Pacific region and weakly negative over the Caribbean. Similarly, trends in heavy and very heavy precipitation were positive over the Indian Ocean, negative over the South Pacific region and close to zero over the Caribbean.

Recent projections of temperature for the Pacific have been undertaken using models from the CMIP3 ensembles (Perkins et al., 2011) that have been assessed in terms of their ability to capture important climate features of the region such as the South Pacific Convergence Zone (Brown et al., 2011) and indicate that, under an A2 emission scenario, most of the South Pacific region will warm by 3°C by 2081-2100 relative to 1981-2000, precipitation will increase by 60% over the Equator, and wind speed will decrease along the Equator and increase further south and tend to a more easterly flow. For the Caribbean, temperatures are projected to increase across the region by 1-4˚C over 2071-2100 relative to 1961-1990 under A2 and B2 scenarios and rainfall is mainly projected to decrease by 25-50% except in the north (Campbell et al., 2011).

Given the low elevation of many small islands, sea level extremes are of particular relevance. In the Pacific, sea level extremes are strongly influenced by tidal extremes (Chowdhury et al., 2007; Merrifield et al., 2007) and depending on whether the tide behaviour is mostly semi-diurnal (two high and low tides per day) or diurnal (one high and low tide per day), there will be a clustering of high spring tides around the time of the equinoxes or the solstices. In addition, ENSO has a strong influence such that sea levels and their extremes are positively (negatively) correlated with the Southern Oscillation Index in the tropical Pacific west (east) of 180° (Church et al., 2006a; Menendez et al., 2010). Tides and ENSO have contributed to the more frequent occurrence of sea level extremes and associated flooding experienced at some Pacific Islands such as Tuvalu in recent years and make the task of determining the relative roles of these natural effects and mean sea level rise difficult (Lowe et al., 2010). Furthermore, the steep shelf margins that surround many islands and atolls in the Pacific support larger wave-induced contributions to sea level anomalies. This suggests that waves are also likely to be a major contributor to positive sea level anomalies for small islands, which has been found to be the case at Midway Atoll in the northern tropical Pacific (Hoeke et al., 2011). Unfortunately, wave observations (including wave direction) are sparse, including those that are co-located with tide gauges, which would facilitate more comprehensive studies of tide, surge and wave extremes in the region (Lowe et al., 2010).

Anthropogenic-induced oceanic changes may reduce the resilience of coral atolls to extreme events such as high waves and storm surge and this may exacerbate extreme impacts. Coral atolls are mainly composed of unlithified or poorly consolidated carbonate sand and gravel, which is supplied by the surrounding reefs. Storms and swell are in many instances the agents of delivery of carbonate material to the shores of atolls (Woodroffe, 2008; Webb and Kench, 2010) and so the health of coral reefs is therefore important for the long term provision of carbonate material for the atolls as sea levels rise in the future. Oceanic changes that could reduce the health of the surrounding reefs and therefore potentially increase severe weather induced erosion and inundation are: (1) warming of the surface ocean, (2) ocean acidification induced by increases in atmospheric carbon dioxide being absorbed into the oceans, and (3) reduction in oxygen concentration in the ocean due to a temperature-driven change in gas solubility. Surface warming of the oceans can itself directly impact biodiversity by slowing or preventing growth in temperature-sensitive species. One of the most well-known biological impacts of warming is coral bleaching, but ocean acidification also plays a role in lowering coral growth rates (Bongaerts et al., 2010). A secondary impact of warming is the potential reduction in oxygen concentrations due to decline in the chemical capacity of seawater to retain dissolved oxygen at higher temperatures (Whitney et al., 2007). It has been predicted that deoxygenation will occur at 1 – 7% over the next century via this mechanism alone, continuing for 1000 years or more into the future (Keeling et al., 2010). An important impact may be an expansion of already existing oxygen minimum zones, especially in tropical oceans. Quantifying these changes and understanding their impact on coral reef health will be important to understanding the impact of anthropogenic climate change.

In summary, the reported increases in warm days and nights and decreases in cool days and nights are of medium confidence over the Caribbean and of low confidence over the Pacific and Indian oceans. There is high confidence in the projected temperature increases across the Pacific and Caribbean. There is insufficient evidence at this time to assess observed trends and future projections in rainfall. The unique situation of small islands in their maritime environments leads to an additional emphasis on oceanic information to understand the risks of climate change. Knowledge and data that is particularly relevant is limited at this stage, and if not

addressed, will hamper efforts to quantify the risks and formulate sound adaptation responses to future climate change for the inhabitants of small islands.

END BOX 3.3 HERE

The AR4, stated with *very high confidence* that the impact of climate change on coasts is exacerbated by increasing human-induced pressures. Consistent with that assessment, the small number of studies that have been completed since the AR4 have been either unable to attribute coastline changes to specific causes in a quantitative way or else find strong evidence for non-climatic causes that are natural and/or anthropogenic.

The AR4 reported with *very high confidence* that coasts will be exposed to increasing risks, including coastal erosion, over coming decades due to climate change and sea level rise, both of which will be exacerbated by increasing human-induced pressures (Nicholls et al., 2007). However it was also noted that since coasts are dynamic systems, adaptation to climate change required understanding of processes operating on decadal to century time scales, yet this understanding was least developed.

Because of the diverse and complex nature of coastal impacts, assessments of the future impacts of climate change have focussed on a wide range of questions and employed a diverse range of methods, making direct comparison of studies difficult (Nicholls, 2010). Two types of studies are reviewed briefly here; the first are assessments, typically undertaken at the country or regional scale and which combine information on physical changes with the socio-economic implications (e.g., Nicholls and de la Vega-Leinert, 2008); the second type are studies oriented around improved scientific understanding of the impacts of climate change. In terms of coastal assessments, Aunan and Romstad (2008) reported that Norway's generally steep and resistant coastlines contribute to a low physical susceptibility to accelerated sea level rise. Nicholls and de la Vega-Leinert (2008) reported that large parts of the coasts in Great Britain (including England, Wales, and Scotland) already experience problems, including sediment starvation and erosion, loss/degradation of coastal ecosystems, and significant exposure to coastal flooding. Lagoons, river deltas and estuaries are assessed as being particularly vulnerable in Poland (Pruszak and Zawadzka, 2008). In Estonia, Kont et al. (2008) reported increased beach erosion, which is believed to be the result of recent increased storminess in the eastern Baltic Sea, combined with a decline in sea-ice cover during the winter. Sterr (2008) reported that for Germany there is a high level of reliance on hard coastal protection against extreme sea level hazards which will increase ecological vulnerability over time. A coastal vulnerability assessment for Australia (Department of Climate Change, 2009), characterised future vulnerability in terms of coastal geomorphology, sediment type and tide and wave characteristics, from which it concluded that the tropical northern coastline would be most sensitive to changes in tropical cyclone behaviour while health of the coral reefs may also influence the tropical eastern coastline. The mid-latitude southern and eastern coastlines were expected to be most sensitive to changes in mean sea level, wave climate and changes in storminess. A comparative study of the impact of sea level rise on coastal inundation across 84 developing countries showed that the greatest vulnerability to a 1 m sea level rise was in East Asia and the Pacific in terms of land area, population, GDP, agricultural, urban and wetland areas, whereas sub-Saharan Africa was least affected (Dasgupta et al., 2009).

New models have been developed for the assessment of coastal vulnerability at the global to national level (Hinkel and Klein, 2009). At the local to regional scale, new techniques and approaches have also been developed to better quantify impacts from inundation due to future sea level rise. Bernier et al. (2007) evaluated species vulnerability to inundation from future sea level rise using seasonal return periods of high water and showed that increased inundation of wetlands from spring time storm surges under sea level rise scenarios could adversely affect bird breeding cycles. McInnes et al. (2009a) developed spatial maps of stormtide and using a simple inundation model with high resolution LiDAR data and a land subdivisions data base, identified the impact of inundation on several coastal towns along the southeastern Australian coastline under future sea level and wind speed scenarios. Probabilistic approaches have also been used to evaluate extreme sea level exceedance under uncertain future sea level rise scenarios. Purvis et al. (2008), assumed a plausible probability distribution to the range of future sea level rise estimates and used Monte-Carlo sampling to apply the sea level change to a two-dimensional coastal inundation model. They showed that by evaluating the possible flood related losses (in monetary terms) in this framework they were able to represent spatially the higher losses associated with the low frequency but high impact events instead of considering only a single midrange scenario. Hunter (2010) presented a method of combining sea-level extremes evaluated from observations with projections of sea level rise to 2100 to evaluate the probabilities of extreme events being exceeded over different future time horizons. There have also been further developments in coastal erosion modelling within probabilistic frameworks that can take into account storm duration and sequencing (i.e., the compound effects on beach erosion that result from storms that occur in short succession), although such methods have not as yet been applied in a climate change context, (Callaghan et al., 2008). Along the Portuguese coast, Andrade et al. (2007) found that projected future climate in the HadCM3 model would not affect wave height along this coastline but the rotation in wave direction will increase the net littoral drift and the erosional response. On the U.K. East Anglia coast, the effect of sea level rise, surge and wave climate change on the inshore wave climate was evaluated and the frequency and height of extreme waves were found to increase in the north of the domain (Chini et al., 2010). On the basis of modelling various climate change scenarios over the next 25 years, Coelho et al. (2009) concluded that the effects of sea level rise are less important than changes in wave action along a stretch of the Portuguese coast. Modelling of the evolution of soft rock shores with rising sea levels has revealed a relatively simple relationship between sea level rise and the equilibrium cliff profile (Walkden and Dickson, 2008).

To summarise, recent observational studies that identify trends and impacts at the coast are low in regional coverage and furthermore the quantity and quality of data and methods to attribute changes to particular causes is also low, which means there is *low confidence* that anthropogenic climate change has been a major cause of the observed changes. However, recent coastal assessments at the national and regional scale and process-based studies have provided further evidence of the vulnerability of low-lying coastlines to rising sea levels and erosion, so that in the absence of adaptation there is *high confidence* that locations currently experiencing adverse impacts such as coastal erosion and inundation will continue to do so in the future.

3.5.6. Glacier, geomorphological and geological impacts

Mountains are prone to mass movements including landslides, avalanches, debris flows and flash floods that can lead to disasters. Changes in mountain glaciers affect these processes, as well as water supply and hydropower generation. Many of the world’s high mountain ranges are situated at the margins of tectonic plates, increasing the possibility of potentially hazardous interactions between climatic and geological processes. The principal drivers are glacier ice mass loss, permafrost degradation, and possible increases in the intensity of precipitation (Liggins et al., 2010; McGuire, 2010). The projected consequences are changes in mass movement on shorter contemporary timescales, and seismicity and volcanic activity on longer, century to millennium timescales. The climate variability that influences these geomorphological phenomenon are simulated by climate models but these phenomena are typically of a regional or local nature. GCMs do not simulate local and regional climate variables with sufficient detail and accuracy to provide high confidence in projections of changes in geomorphological phenomena.

The AR4 assessed past changes in glaciers and concluded that the widespread retreat of glaciers in the world is consistent with warming. However, the impacts of glacier retreat on the natural physical system in the context of changes in extreme events were not assessed in detail. Additionally, the AR4 did not assess geomorphological and geological impacts that might result from anthropogenic climate change.

The most studied change in the high-mountain environment has been the retreat of glaciers (Paul et al., 2004; Kaser et al., 2006; Larsen et al., 2007; Rosenzweig et al., 2007). Alpine glaciers around the world achieved their maximum extent at the end of the Little Ice Age (~1850), and have retreated since then (Oerlemans, 2005), with an accelerated decay during the past several decades (Zemp et al., 2007). Glaciers have retreated in many parts of the world (Francou et al., 2000; Cullen et al., 2006; Thompson et al., 2006; Larsen et al., 2007; Schiefer et al., 2007; Paul and Haeberli, 2008). Rates of retreat that exceed historical experience and internal (natural) variability have become apparent since the beginning of the 21st century (Reichert et al., 2002; Haeberli and Hohmann, 2008).

Outburst floods from lakes dammed by glaciers or unstable moraines (or “Glacial lake outburst floods”, GLOFs) are commonly a result of glacier retreat and formation of lakes behind unstable natural dams (Clarke, 1982; Clague and Evans, 2000; Huggel et al., 2004; Dussaillant et al.2010). In the past century GLOFs have caused disasters in many high-mountain regions of the world (Rosenzweig et al., 2007), including the Andes (Reynolds et al., 1998; Carey, 2005; Hegglin and Huggel, 2008), the Caucasus and Central Asia (Narama et al., 2006; Aizen et al., 2007), the Himalayas (Vuichard and Zimmermann, 1987; Richardson and Reynolds, 2000; Xin et al., 2008; Bajracharya and Mool, 2009; Osti and Egashira, 2009), North America (Clague and Evans, 2000; Kershaw et al., 2005), and the European Alps (Haeberli, 1983; Haeberli et al., 2001; Vincent et al., 2010). However, because GLOFs are relatively rare, it is unclear whether their occurrence is changing on either the regional or global scale. Clague and Evans (2000) argue that outburst floods from moraine-dammed lakes may have peaked due to a reduction in the number of the lakes since the end of the Little Ice Age. In contrast, a small, but not statistically significant increase of GLOF events was observed in the Himalayas over the period 1940-2000 (Richardson and Reynolds, 2000) though there has been no GLOF since the 1998 event of Tam Pokhari (Osti and Egashira, 2009) in the region.

Evidence of degradation of mountain permafrost and attendant slope instability has emerged from recent studies in the European Alps (Gruber and Haeberli, 2007; Huggel, 2009) and other mountain regions (Niu et al., 2005; Geertsema et al., 2006; Allen et al., 2011). This evidence includes several recent rock falls, rock slides, and rock avalanches in areas where permafrost thaw is occurring. Landslides with volumes ranging up to a few million cubic metres occurred in the Mont Blanc region (Barla et al., 2000), in Italy (Sosio et al., 2008; Huggel, 2009; Fischer et al., 2011a), in Switzerland and in British Columbia (Evans and Clague, 1998; Geertsema et al., 2006). Very large rock and ice avalanches with volumes of 30 to over 100 million m^3 include the 2002 Kolka avalanche in the Caucasus (Haeberli et al., 2004; Kotlyakov et al., 2004; Huggel et al., 2005), the 2005 Mt. Steller rock avalanche in the Alaska Range (Huggel et al., 2008), the 2007 Mt. Steele ice and rock avalanche in the St. Elias Mountains, Yukon (Lipovsky et al., 2008), and the 2010 Mt. Meager rock avalanche and debris flow in the Coast Mountains of British Columbia.

Quantification of possible trends in the frequency of landslides and ice avalanches in mountains is difficult due to incomplete documentation of past events, especially those that happened before satellite observations became available. Nevertheless, there has been an apparent increase in large rock slides during the past two decades, and especially during the first years of the 21st century in the European Alps, the Southern Alps of New Zealand (Allen et al., 2011; Fischer et al., 2011b) and in northern British Columbia (Geertsema et al., 2006) in tandem with temperature increases, glacier shrinkage, and permafrost degradation.

Research, however, has not yet provided any clear indication of a change in the frequency of debris flows due to recent deglaciation. Debris flow activity at a local site in the Swiss Alps was higher during the 19th century than today (Stoffel et al., 2005). In the French Alps no significant change in debris flow frequency has been observed since the 1950s in terrain above elevations of 2200 m (Jomelli et al., 2004). Processes not, or not directly, driven by climate, such as sediment yield can also be important for changes in the magnitude or frequency alpine debris flows (Lugon and Stoffel, 2010).

Debris flows from both glaciated and unglaciated volcanoes, termed lahars, can be particularly large and hazardous. Lahars produced by volcanic eruptions on the glacier-clad Nevado del Huila volcano in Colombia in 2007 and 2008 were the largest, rapid mass flows on Earth in recent years. Similarly, large mass flows occur on ice-covered active volcanoes in Iceland (Björnsson, 2003), including Eyjafjallajökull in 2010. Large rock and ice avalanches, with volumes up to 30 million m^3, have happened frequently (averaged about one every 4 years) on the glaciated Alaskan volcano, Iliamna, are thought to be related to elevated volcanic heat flow and possibly meteorological conditions (Huggel et al., 2007). In 1998, intense rainfall mobilised pyroclastic material on the flanks of Vesuvius and Campi Flegrei volcanoes, feeding *ca.* 150 debris flows that damaged nearby communities and resulted in 160 fatalities (Bondi and Salvatori, 2003). In the same year, intense precipitation associated with Hurricane Mitch triggered a small flank collapse at Casita volcano in Nicaragua. This slope failure transformed into debris flows that destroyed two towns and claimed 2,500 lives (Scott et al., 2005). Glacier retreat in the area of Bering Glacier in southeast Alaska appears to be modulating the recent seismic record (Sauber et al., 2000; Doser et al., 2007; Sauber and Ruppert, 2008) and may have been a contributing factor for the 1972 St. Elias earthquake (Sauber and Molnia, 2004).

A variety of climate and weather events can have geomorphological and geological impacts. Warming and degradation of permafrost affect slope stability. For example, the 2003 European summer heat wave (Section 3.3.1) caused rapid thaw and thickening of the active layer and trigging a large number of mainly small rock falls (Gruber et al., 2004; Gruber and Haeberli, 2007). Permafrost thaw may increase both the frequency and magnitude of debris flows (Zimmermann et al., 1997; Rist and Phillips, 2005). The frost table at the base of the active layer is a barrier to groundwater infiltration and can cause the overlying non-frozen sediment to become saturated. Snow cover can also affect debris flow activity by supplying additional water to the soil, increasing pore water pressure and initiating slope failure (Kim et al., 2004). Many of the large debris flows in the Alps in the past 20 years were triggered by intense rainfall in summer or fall when the snowline was elevated (Rickenmann and Zimmermann, 1993; Chiarle et al., 2007). Warming may increase the flow speed of frozen bodies of sediment (Kääb et al., 2007; Delaloye et al., 2008; Roer et al., 2008). Rock slopes can fail after they have been steepened by glacial erosion or unloaded (debuttressed) following glacier retreat (Augustinus, 1995). Although it may take centuries or even longer for a slope to fail following glacier retreat, recent landslides demonstrate that some slopes can respond to glacier downwasting within a few decades or shorter (Oppikofer et al., 2008). Twentieth-century warming may have penetrated some decametres into thawing steep rock slopes inn high mountains (Haeberli et al., 1997). Case studies indicate that both small and large slope failures can be triggered by exceptionally warm periods of weeks to months (Gruber et al., 2004; Huggel, 2009; Fischer et al., 2011a).

The spatial and temporal patterns of precipitation, the intensity and duration of rainfall, and antecedent rainfall are important factors in triggering shallow landslides (Iverson, 2000; Wieczorek et al., 2005; Sidle and Ochiai, 2006). In some regions antecedent rainfall is probably a more important factor than rainfall intensity (Kim et al., 1991; Glade, 1998), whereas in other regions rainfall duration and intensity are the critical factors (Jakob and Weatherly, 2003). Landslides in temperate and tropical mountains that have no seasonal snow cover are not temperature-sensitive and may be more strongly influenced by human activities such as poor land-use practises, deforestation, and overgrazing (Sidle and Ochiai, 2006).

Rock and ice avalanches on glaciated volcanoes can be triggered by heat generated by volcanic activity. Their incidence may increase with rising air and rock temperatures (Gruber and Haeberli, 2007) or during or following brief, anomalously warm events (Huggel et al., 2010). Landslides are also favoured by glacier flows, which may destabilize or oversteepen slopes (Tuffen, 2010); by melting ice, which may create weak zones at ice-bedrock interfaces (Huggel, 2009), and by shallow hydrothermal alteration driven by snow and ice melt (Huggel, 2009). On unglaciated high volcanoes in the Caribbean, Central America, Europe, Indonesia, the Philippines and Japan, an increase in total rainfall or an increase in the frequency or magnitude of severe rainstorms could cause more frequent debris flows by mobilizing unconsolidated, volcanic regolith and by raising pore-water pressures, which could lead to deep-seated slope failure. Heavy rainfall events could also influence the behaviour of active volcanoes. For example, Mastin (1994) attributes the violent venting of volcanic gases at Mount St Helens between 1989 and 1991 to slope instability or accelerated growth of cooling fractures within the lava dome following rainstorms, and Matthews et al. (2002) link episodes of intense tropical rainfall with collapses of the Soufriere Hills lava dome on Montserrat in the Caribbean. A large reduction in glacier cover may be responsible for an increase in seismicity in southeast Alaska where earthquake faults are at the threshold of failure (Sauber and Molnia, 2004; Doser et al., 2007). An increase in the frequency of small earthquakes in the Icy Bay area, also in southeast Alaska, is interpreted to be a crustal response to a glacier wastage between 2002 and 2006 (Sauber and Ruppert, 2008). Large-scale ice-mass loss in glaciated volcanic terrain reduces the load on the crust and uppermost mantle, facilitating of the rise of more magma into the crust (Jull and McKenzie, 1996) and allowing magma to reach the surface more easily (Sigmundsson et al., 2010). At the end of the last glaciation, this mechanism resulted in a more than 10-fold increase in the frequency of volcanic eruptions in Iceland (Sinton et al., 2005). Widespread uplift of up to 20 mm y^{-1} is currently occurring in response to thinning of Vatnajökull Ice Cap in Iceland and is expected to cause a future increase in volcanic activity (Sigmundsson et al., 2010).

Phenomena such as ice avalanches, GLOFs, and some landslides that are related to glacier retreat are temperature-sensitive and thus are *likely* to have been affected by anthropogenic warming. There is *low confidence* in an anthropogenic influence on precipitation-driven phenomena such as shallow landslides because an anthropogenic influence on regional precipitation has yet to be established with confidence. Poor land-use practices also may contribute to landslides, and such factors complicate the attribution of changes in geomorphogical and geological impacts to a single factor (e.g., Sidle and Ochiai, 2006).

The AR4 projected that glaciers in mountains will lose additional mass over this century because more ice will be lost due to summer melting than is replenished by winter precipitation (Meehl et al., 2007b). The total area of glaciers in the European Alps may decrease by 20% to more than 50% by 2050 (Zemp et al., 2006; Huss et al., 2008). Atmospheric warming favours rapid glacier mass loss and related mass movements (Huggel et al., 2011). The projected glacier retreat in the 21st century will *likely* form new, potentially unstable lakes. Probable sites of new lakes have been identified for some alpine glaciers (Frey et al., 2010). Rock slope and moraine failures may trigger damaging surge waves and outburst floods from these lakes. The temperature rise also will result in gradual permafrost degradation (Haeberli and Burn, 2002; Harris et al., 2009). Warm permafrost (mean annual rock temperature ~- 2 to 0°C), which is more susceptible to slope failures than cold permafrost, may rise in elevation a few hundred metres during the next 100 years (Noetzli and Gruber, 2009). The response of bedrock temperatures to surface warming through thermal conduction will be slow, but warming will eventually penetrate to considerable depths in steep rock slopes (Noetzli et al., 2007). Other heat transport processes such as advection, however, may induce warming of bedrock at much faster rates (Gruber and Haeberli, 2007). The response of firn and ice temperatures to an increase in air temperature increase is faster and non-linear (Haeberli and Funk, 1991; Suter et al., 2001; Vincent et al., 2007). Latent heat effects from refreezing melt water can amplify the increase in air temperature in firn and ice (Huggel, 2009; Hoelzle et al., 2010). At higher temperatures, more ice melts and the strength of the remaining ice is lower; as a result, the frequency and perhaps size of ice avalanches may increase (Huggel et al., 2004; Caplan-Auerbach and Huggel, 2007). Warm extremes can trigger large rock and ice avalanches (Huggel et al., 2010), and warm extremes have been projected to increase by several fold by 2050 (Huggel et al., 2010).

Current low levels of seismicity in Antarctica and Greenland may be a consequence of ice-sheet loading. Isostatic rebound associated with accelerated deglaciation of these regions may result in an increase in earthquake activity, perhaps on a timescales as short as 10 – 100 years (Turpeinen et al., 2008; Hampel et al., 2010). Future ice-mass loss on glaciated volcanoes, notably in Iceland, Alaska, Kamchatka, the Cascade Range in the northwest USA, and the Andes, could lead to eruptions, either as a consequence of reduced load pressures on magma chambers or through increased magma–water interaction. Reduced ice load arising from future thinning of Iceland's Vatnajökull Ice Cap is projected to result in an additional 1.4 km^3 of magma produced in the underlying mantle every century (Pagli and Sigmundsson, 2008). Ice-unloading will also promote failure of shallow magma reservoirs; the most likely consequence being a small perturbation of the natural eruptive cycle (Sigmundsson et al., 2010). Ice thinning of 100 m or more on volcanoes with glaciers more than 150 m thick, such as Sollipulli in Chile, may cause more explosive eruptions, with increased tephra hazards (Tuffen, 2010). Additionally, the potential for edifice lateral collapse could be enhanced due to loss of support previously provided by ice (Tuffen, 2010) or to elevated pore-water pressures arising from meltwater (Capra, 2006; Deeming et al., 2010). The likelihood of both volcanic and non-volcanic landslides may also be increased due to greater availability of water, which could destabilize slopes. Many volcanoes provide a ready source of unconsolidated debris that can be rapidly transformed into potentially hazardous lahars by extreme precipitation events. Volcanoes in coastal, near-coastal or island locations in the tropics are particularly susceptible to torrential rainfall associated with tropical cyclones, and such events are projected to increase through this century (see section 3.4.5). The impact of future, large, explosive, volcanic eruptions may also be exacerbated by an increase in extreme precipitation events, by providing an effective means of transferring large volumes of unconsolidated ash and pyroclastic flow debris from the flanks of volcanoes into downstream areas. Following the 1991 Pinatubo eruption in the Philippines, heavy rains associated with tropical storms moved large volumes of volcanic sediment. The sediment dammed rivers, causing massive flooding across the region that continued for several years after the eruption ended (Newhall and Punongbayan, 1996).

In summary, many weather and climate events can have geomorphological and geological impacts. Many other factors, however, also influence these impacts, thus it is difficult to attribute any recent trends in impacts to climate change. As well, the availability of representative long time series of such impacts are rare, so it is difficult to quantitatively identify long-term trends. Nevertheless, there is *medium confidence* that high-mountain debris flows will begin earlier in the year because of earlier snow melt. There is *low confidence* in projected changes in the magnitude and frequency of shallow landslides in temperate and tropical regions, as they depend mainly on frequency and intensities of rainfall events and anthropogenic land-use. It is *likely* that continued permafrost degradation and glacier retreat will further decrease the stability of rock slopes, although there is *low confidence* regarding future locations and timing of large rock avalanches, as these depend on local geological conditions and other non-climatic factors.

3.5.7. High-Latitude Changes including Permafrost

Permafrost is widespread in Arctic, Subarctic, in ice-free areas of Antarctica, and in high-mountain regions, and permafrost regions occupy approximately 23 million km^2 of land areas in the Northern Hemisphere (Zhang et al., 1999). Melting of massive ground ice and thawing of ice-rich permafrost can lead to subsidence of ground surface and to the formation of uneven topography known as thermokarst, generating dramatic changes in ecosystems, landscapes, and infrastructure performance (Nelson et al., 2001; Walsh, 2005). The active layer (near surface layer that thaws and freezes seasonally over permafrost) plays an important role in cold regions because most ecological, hydrological, biogeochemical and pedogenic (soil-forming) activity takes place within it (Hinzman et al., 2005).

Limited observations show that temperatures at the top of the permafrost have increased by up to 3°C since the early 1980s (Lemke et al., 2007; Harris et al., 2009). Over the high Arctic such as in northern Alaska (Osterkamp, 2005, 2007) and Russia (Obserman and Mazhitova, 2001), permafrost temperatures have increased by about 2 to 3°C since the mid-1980s. The magnitude of permafrost temperature increase is up to 1.0°C in the Interior of Alaska (Osterkamp, 2005, 2007), much of the Canadian Arctic (Smith et al., 2005b), Mongolia (Sharkhuu, 2003), and on the Tibetan Plateau since the 1980s (Cheng and Wu, 2007). Generally speaking, the magnitude of permafrost temperature increase in continuous permafrost regions is greater than in discontinuous permafrost regions (Osterkamp, 2007). Increases in snow depth may contribute significantly to the greater permafrost temperature increase in the high Arctic, and contribute to local and regional variability of permafrost temperature increase (Zhang et al., 2005). When the other conditions remain constant, active layer thickness is expected to increase in response to climate warming, especially in summer. Observations show that active layer thickness has increased about 20 cm in the Russian Arctic between the early 1960s to 2000 (Zhang et al., 2005), up to 1.0 m over the Qinghai-Tibetan Plateau since the early 1980s (Wu and Zhang, 2010), with no significant trend in North American Arctic since the early 1990s (Brown et al., 2000). Extensive thermokarst development has been found in Alaska (Yoshikawa and Hinzman, 2003; Osterkamp et al., 2009), in central Yakutia (Gavriliev and Efremov, 2003), and on the Qinghai-Tibetan Plateau (Niu et al., 2005). Significant expansion and deepening of thermokarst lakes were observed near Yakutsk with subsidence rates of 17 to 24 cm yr^{-1} from 1992–2001 (Fedorov and Konstantinov, 2003). Satellite remote sensing data show that thaw lake surface area has increased in continuous permafrost regions and decreased in discontinuous permafrost regions (Smith et al., 2005a). Coasts with ice-bearing permafrost that are exposed to the Arctic Ocean are very sensitive to permafrost degradation. Some Arctic coasts are retreating at a rapid rate of 2 to 3 m yr^{-1} and the rate of erosion along Alaska's northeastern coastline has doubled over the past 50 years (Karl et al., 2009).

Increases in air temperature are in part responsible for the observed increase in permafrost temperature over the Arctic and Subarctic, and changes in snow cover also play a critical role (Osterkamp, 2005; Zhang et al., 2005). Trends towards earlier snowfall in autumn and thicker snow cover during winter have resulted in stronger snow insulation effect, and as a result a much warmer permafrost temperature than air temperature in the Arctic. The lengthening of the thaw season and increases in summer air temperature have resulted in changes in active layer thickness. A model simulation suggests there will only be about 10% of current near-surface permafrost remaining by 2100 (Lawrence and Slater, 2005). The combination of Arctic sea ice retreat, storm activity increase, and permafrost degradation is responsible for rapid Arctic coast erosion in recent decades (Atkinson et al., 2006). Expansion of lakes in the continuous permafrost zone may be due to thawing of ice-rich permafrost and melting of massive ground ice, while decreases in lake area in the discontinuous permafrost zone may be due to lake bottom drainage (Smith et al., 2005a). Overall, increased air temperature over high latitudes is primarily responsible for development of thermokarst terrains and thaw lakes.

In summary, it is *likely* that there has been an increased thawing of permafrost in recent decades, and it is *likely* that it has had physical impacts. It is *very likely* that permafrost temperatures will continue to increase, and it is *likely* that there will be widespread increases in active layer thickness and large reduction in the area of permafrost in the Arctic and Subarctic. Due to sea ice retreat, and permafrost degradation, with possibly a

contribution from more storminess, the frequency and magnitude of the rate of Arctic coastal erosion will *likely* increase.

3.5.8. Sand and Dust Storms

Sand and dust storms are widespread natural phenomena in many parts of the world. Heavy dust storms disrupt human activities. Dust aerosols in the atmosphere can cause a suite of health impacts including respiratory problems (Small et al., 2001). The long-range transport of dust can affect conditions at long distances from the dust sources, linking the biogeochemical cycles of land, atmosphere and ocean (Martin and Gordon, 1988; Bergametti and Dulac, 1998; Kellogg and Griffin, 2006). For example, dust from the Saharan region and from Asia may reach North America (McKendry et al., 2007). Most of the GCMs have implemented the processes to simulate sand and dust storms (Textor et al., 2006). Climate variables that are most important to dust emission and transportation such as soil moisture, precipitation and wind are still subject to large uncertainties in the simulations. As a result, the sand and dust storm simulations have large uncertainties as well.

The Sahara (especially Bodélé Depression in Chad) and east Asia have been recognized as the strongest dust sources globally (Goudie, 2009). Over the past few decades, the frequency of dust events has increased in some regions such as the Sahel zone of Africa (Goudie and Middleton, 1992), and decreased in some other regions such as China (Zhang et al., 2003), but there seems to also be an increase in more recent years (Shao and Dong, 2006). Despite the importance of African dust, studies on long-term change in Sahel dust are limited. However, dust transported far away from the source region may provide some evidence of long-term changes in the Sahel region. The African dust transported to Barbados began to increase in the late 1960s and through the 1970s; transported dust reached a peak in the early 1980s but remains high into the present (Prospero and Lamb, 2003; Prospero et al., 2009).

Surface soil dust concentration during a sand and dust storm is controlled by a number of factors. The driving force for the production of dust storms is the surface wind associated with cold frontal systems sweeping across the dry desert areas and lifting soil particles in the atmosphere. Dust emissions are also controlled by the surface conditions in source regions such as the desert coverage distributions, snow cover and soil moisture. In the Sahel region, the elevated high level of dust emission is related to the persistent drought since the 1970s, and to long-term changes in the North Atlantic Oscillation (Ginoux et al., 2004; Chiapello et al., 2005; Engelstaedter et al., 2006), and perhaps to North Atlantic SST as well (Wong et al., 2008). The desert areas increased by ~2 to ~7% (Zhong, 1999) in China during 1960-2000, when the dust storm frequency decreased. A 44-year simulation study of Asian soil dust production with a dynamic desert distribution from 1960 to 2003 suggests that climatic variations have played a major role in the declining trends in dust emission and storm frequencies (Zhang et al., 2003; Zhou and Zhang, 2003; Zhao et al., 2004) in China (Gong et al., 2006). Changes in wind (Wang et al., 2006b), meridional temperature gradients and cyclone frequencies (Qian et al., 2002), large-scale circulations such as the Asian polar vortex (Gong et al., 2006), the Siberia high (Ding et al., 2010), rainfall and vegetation (Zhou and Zhang, 2003) all contributed to the decrease in the observed dust frequency in China. Overall, the observed changes in dust activity are mainly the result of long-term changes in the climate, such as wind and moisture conditions in the dust source regions. Changes in large-scale circulation play an additional role in the long-distance transport of dust. However, understanding of the physical mechanisms of the long-term trends in dust activity is not complete; for example, there are a large number of potential factors affecting dust frequency in China, but their relative importance is uncertain.

Future dust activity depends on two main factors: land use in the dust source regions, and climate both in the dust source region and large-scale circulation that affects long distance dust transport. Studies on projected future dust activity are very limited. It is difficult to project future land use. Precipitation, soil moisture, and runoff, have been projected to decrease in major dust source regions (Figure 10.12, Meehl et al., 2007b). Thomas et al. (2005) suggest that dune fields in southern Africa can be reactivated, and sand will become significantly exposed and move, as a consequence of 21st century warming. A study based on simulations from two climate models also suggests increased desertification in arid and semi-arid China, especially in the second half of the 21st century (Wang et al., 2009e). However, projected changes in wind are lacking.

In summary, because there is high uncertainty in simulating sand and dust storms including important climate variables such as soil moisture, precipitation, and wind that affect dust storms (and because land-use changes and other non-climate factors influence dust storms substantially), there is *low confidence* in projecting future dust storm changes.

START FAQ 3.1 HERE

FAQ 3.1: Is the Climate Becoming More Extreme?

While there is evidence that increases in greenhouse gases have likely caused changes in some types of extremes, there is no simple answer to the question of whether the climate, in general, has become more or less extreme. Both the terms "more extreme" and "less extreme" can be defined in different ways, resulting in different characterizations of observed changes in extremes. Additionally, from a physical climate science perspective it is difficult to devise a comprehensive metric that encompasses all aspects of extreme behaviour in the climate.

One approach for evaluating whether the climate is becoming more extreme would be to determine whether there have been changes in the typical range of variation of specific climate variables. For example, if there was evidence that temperature variations in a given region had become significantly larger than in the past, then it would be reasonable to conclude that temperatures in that region had become more extreme. More simply, temperature variations might be considered to be becoming more extreme if the difference between the highest and the lowest temperature observed in a year is increasing. According to this approach, daily temperature over the globe may have become less extreme because there have generally been greater increases in annual minimum temperatures globally than in annual maximum temperatures, over the second half of the 20th century. On the other hand, one might conclude that daily precipitation has become more extreme because observations suggest that the magnitude of the heaviest precipitation events has increased in many parts of the world. Another approach would be to ask whether there have been significant changes in the frequency with which climate variables cross fixed thresholds that have been associated with human or other impacts. For example, an increase in the mean temperature usually results in an increase in hot extremes and a decrease in cold extremes. Such a shift in the temperature distribution would not increase the "extremeness" of day-to-day variations in temperature, but would be perceived as resulting in a more extreme warm temperature climate, and a less extreme cold temperature climate. So the answer to the question posed here would depend on the variable of interest, and on which specific measure of the extremeness of that variable is examined. As well, to provide a complete answer to the above question, one would also have to collate not just trends in single variables, but also indicators of change in complex extreme events resulting from a sequence of individual events, or the simultaneous occurrence of different types of extremes. So it would be difficult to comprehensively describe the full suite of phenomena of concern, or to find a way to synthesize all such indicators into a single extremeness metric that could be used to comprehensively assess whether the climate as a whole has become more extreme from a physical perspective. And to make such a metric useful to more than a specific location, one would have to combine the results at many locations, each with a different perspective on what is "extreme".

Three types of metrics have been considered to avoid these problems, and thereby allow an answer to this question. One approach is to count the number of record-breaking events in a variable and to examine such a count for any trend. However, one would still face the problem of what to do if, for instance, hot extremes are setting new records, while cold extremes were not occurring as frequently as in the past. In such a case counting the number of records might not indicate whether the climate was becoming more or less extreme, rather just whether there was a shift in the mean climate. Also, the question of how to combine the numbers of record-breaking events in various extremes (e.g., daily precipitation and hot temperatures) would need to be considered. Another approach is to combine indicators of a selection of important extremes into a single index, such as the Climate Extremes Index (CEI) which measures the fraction of the area of a region or country experiencing extremes in monthly mean surface temperature, daily precipitation, and drought. The CEI, however, omits many important extremes such as tropical cyclones and tornadoes, and could, therefore, not be considered a complete index of "extremeness". Nor does it take into account complex or multiple extremes, nor the varying thresholds that relate extremes to impacts in various sectors.

A third approach to solving this dilemma arises from the fact that extremes often have deleterious economic consequences. It may therefore be possible to measure the integrated economic effects of the occurrence of different types of extremes into a common instrument such as insurance payout to determine if there has been an increase or decrease in that instrument. This approach would have the value that it clearly takes into account those extremes with economic consequences. But trends in such an instrument will be dominated by changes in vulnerability and exposure and it will be difficult, if not impossible, to disentangle changes in the instrument caused by non-climatic changes in vulnerability or exposure in order to leave a residual that reflects only changes in climate extremes. For example, coastal development can increase the exposure of populations to hurricanes; therefore, an increase in damage in coastal regions caused by hurricane landfalls will largely reflect changes in exposure and may not be indicative of increased hurricane activity. Moreover, it may not always be possible to associate impacts such as the loss of human life or damage to an ecosystem due to climate extremes to a measurable instrument.

None of the above instruments has yet been developed sufficiently as to allow us to confidently answer the question posed here. Thus we are restricted to questions about whether specific extremes are becoming more or less common, and our confidence in the answers to such questions, including the direction and magnitude of changes in specific extremes, depends on the type of extreme, as well as on the region and season, linked with the level of understanding of the underlying processes and the reliability of their simulation in models.

END FAQ 3.1 HERE

START FAQ 3.2 HERE

FAQ 3.2: Can we Determine Whether Climate Change has Affected Individual Extreme Events?

A changing climate can be expected to lead to changes in climate and weather extremes. But it is challenging to associate a single extreme event with a specific cause such as increasing greenhouse gases because a wide range of extreme events could occur even in an unchanging climate, and because extreme events are usually caused by a combination of factors. Despite this, it may be possible to make an attribution statement about a specific weather event by attributing the changed probability of its occurrence to a particular cause. For instance, the likelihood of heatwaves has increased due to greenhouse warming, while the likelihood of frost or extremely cold nights, has decreased. For example, it has been estimated that human influences have more than doubled the probability of a very hot European summer like that of 2003.

Recent years have seen many extreme events including the extremely hot summer in Europe in 2003 and the intense North Atlantic hurricane seasons of 2004 and 2005. Can the increased atmospheric concentrations of greenhouse gases be considered the 'cause' of such extreme events? That is, could we say these events would NOT have occurred if CO_2 had remained at pre-industrial concentrations? FAQ 3.2, Figure 1 shows the distribution of monthly mean November temperatures averaged across the State of New South Wales in Australia, using data from 1950-2009. The mean temperature for November 2009 (the bar on the far right hand end of the Figure) lies about 3.5 standard deviations above the 1950-2008 mean suggesting that the chance of such a temperature occurring in the 1950-2008 climate (assuming a stationary climate) is quite low. Is this event, therefore, an indication of a changing climate? In the CRUTEM3V global land surface temperature data set, about one in every 900 monthly mean temperatures observed between 1900 and 1949 lies more than 3.5 standard deviations above the corresponding monthly mean temperature for 1950-2008[1]. Since global temperature was lower in the first half of the 20th century, this clearly indicates that an extreme warm event as rare as the 2009 November temperature in New South Wales could have occurred before the effects of greenhouse gas increases were much less pronounced.

A second complicating issue is that extreme events usually result from a combination of factors, and this will make it difficult to attribute an extreme to a single causal factor. So the hot 2003 European, was associated with a persistent high-pressure system (which led to clear skies and thus more solar energy received at the surface) and to dry soil (which meant that less solar energy was used for evaporation, leaving more energy to heat the soil). Another example is that hurricane genesis requires weak vertical wind shear, as well as very warm sea surface temperatures. Since some factors, but not others, may be affected by a specific cause such as increasing greenhouse gas concentrations, it is difficult to separate the human influence on a single, specific extreme event, from other factors influencing the extreme.

However, climate models can sometimes be used to identify if specific factors are changing the likelihood of the occurrence of extreme events. In the case of the 2003 European heat wave, a model experiment indicated that human influences more than doubled the likelihood of having a summer in Europe as hot as that of 2003, as discussed in AR4. The value of such a probability-based approach – 'Does human influence change the likelihood of an event?' – is that it can be used to estimate the influence of external factors, such as increases in greenhouse gases, on the frequency of specific types of events, such as heatwaves or cold extremes. The same likelihood-based approach has been used to examine anthropogenic greenhouse gas contribution to flood probability.

[INSERT FAQ 3.2, FIGURE 1 HERE
FAQ 3.2, Figure 1: The distribution of monthly mean November temperatures averaged across the State of New South Wales in Australia, using data from 1950–2009. Data from Australian Bureau of Meteorology. The mean temperature for November 2009 (the bar on the far right hand end of the Figure) was more than three standard deviations from the long-term mean (calculated from 1950–2008 data).]

[1] We used the CRUTEM3V land surface temperature data. We limit our calculation to grid points with long-term observations, requiring at least 50 non-missing values during 1950-2008 for a calendar month and a grid point to be included. A standard deviation is computed for the period 1950-2008. We then count the number of occurrences when the temperature anomaly during 1900-1949 relative to 1950-2008 mean is greater than 3.5 standard deviations, and compare it with the total number of observations for the grid and month in that period. The ratio between these two numbers is 0.00107.

The discussion above relates to an individual, specific occurrence of an extreme event (e.g., a single heatwave). For the reasons outlined above it remains very difficult to attribute any individual event to greenhouse gas induced warming (even if physical reasoning or model experiments suggest such an extreme may be more likely in a changed climate). However, a long-term trend in an extreme (e.g., heatwave occurrences) is a different matter. It is certainly feasible to test whether such a trend is likely to have resulted from anthropogenic influences on the climate, just as a global warming trend can be assessed to determine its likely cause.

END FAQ 3.2 HERE

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Chapter 3: Changes in Climate Extremes and their Impacts on the Natural Physical Environment

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Tables and Figures

Table 3.1: Overview of considered extremes and summary of observed and projected changes on global scale. Regional details on observed and projected changes in temperature and precipitation extremes are provided in Tables 3.2 and 3.3.

		Observed Changes (since 1950)	Attribution of Observed Changes	Projected Changes (up to 2100)
Weather and climate elements	Temperature (Section 3.3.1)	*Very likely* decrease in number of unusually cold days and nights on the global scale. *Very likely* increase in number of unusually warm days and nights on the global scale. *Likely* increase in warm spells, including heatwaves, in most regions. *Low or medium confidence* in trends in some subregions due either to lack of observations or varying signal within subregions. [Regional details in Table 3.2]	*Likely* anthropogenic influence on global trends in extreme temperature. No attribution of trends on regional scale with a few exceptions.	*Virtually certain* decrease in number of unusually cold days and nights (as defined with1961-1990 climate) on global scale. *Virtually certain* increase in number of unusually warm days and nights on global scale. *Very likely* increase in length, frequency, and/or intensity of warm spells, including heatwaves over most land areas. [Regional details in Table 3.3]
	Precipitation (Section 3.3.2)	*Likely* statistically significant increases in the number of heavy precipitation events (e.g., 95th percentile) in more regions than with statistically significant decreases, but strong regional and subregional variations in the trends. [Regional details in Table 3.2]	*Medium confidence* that changes in extreme precipitation at global scale may have been anthropogenically related.	*Likely* increase in frequency of heavy precipitation events (or increase in proportion of total rainfall from heavy falls) over many areas of the globe, in particular in the high latitudes and tropical regions, and in winter in the northern mid latitudes. [Regional details in Table 3.3]
	Winds (Section 3.3.3)	*Low confidence* in trends because of insufficient evidence	*Low confidence* in the causes of trends	*Low confidence* in projections of extreme winds (with the exception of tropical cyclones)
Phenomena related to weather and climate extremes	Monsoons (Section 3.4.1)	*Low confidence* in trends because of insufficient evidence	*Low confidence* due to *insufficient evidence*	*Low confidence* in projected changes of monsoons, because of lack of consensus between climate models
	El Niño and other modes of variability (Section 3.4.2 and 3.4.3)	*Medium confidence* of past trends towards more frequent central equatorial Pacific El Niño Southern Oscillation (ENSO) events. *Insufficient evidence* for more specific statements on ENSO trends. *Likely* trends in North Atlantic Oscillation (NAO) and Southern Annular Mode (SAM).	*Likely* anthropogenic influence on identified trends in NAO and SAM.	*Low confidence* in projections of changes in behaviour of ENSO and other modes of variability because of insufficient congruence of model projections.
	Tropical cyclones (Section 3.4.4)	*Low confidence* of any robust long-term increases in tropical cyclone activity, after accounting for changes in observing capabilities.	*Low confidence* in attribution of changes in tropical cyclone activity to anthropogenic influences.	*Unlikely* increase in global frequency of tropical cyclones (*likely* decrease or no change). *Likely* increase in mean maximum wind speed, but possibly not in all basins. *Likely* increase in tropical cyclone-related rainfall rates.
	Extra-tropical cyclones (Section 3.4.5)	*Likely* poleward shift in extratropical cyclones. *Low confidence* in regional changes in intensity.	*About as likely as not* anthropogenic influence on poleward shift.	*Likely* impacts on regional cyclone activity but *low confidence* in detailed regional projections. A reduction in the numbers of of mid-latitude storms *is as likely as not*.

				Medium confidence in projected poleward shift of mid-latitude storm tracks.
Impacts on physical environment	Droughts (Section 3.5.1)	*Medium confidence* that some regions of the world have experienced more intense and longer droughts, in particular in southern Europe and West Africa, but also opposite trends exist. [Regional details in Table 3.2]	*Medium confidence* that anthropogenic influence has contributed to the observed increases in droughts	*Medium confidence* in projected increase of duration and intensity of soil moisture and hydrological drought in some regions of the world, in particular in the Mediterranean, Central North America, Southern Mexico and Southern Africa. [Regional details in Table 3.3]
	Floods (Section 3.5.2)	*Low confidence* in changes in the magnitude and frequency in floods at the global level. Earlier occurrence of spring peak river flows in snowmelt- and glacier-fed rivers (*high confidence*),	*Low confidence* that anthropogenic warming has affected the magnitude or frequency of floods, anthropogenic influence on earlier spring peak in snow-dominated regions.	*Low confidence* in global projections of changes in flood magnitude and frequency because of insufficient literature and poor agreement between models. Increase in magnitude and/or frequency anticipated in regions where rainfall extremes are projected to increase. *Very likely* earlier spring peak flows in snowmelt and glacier-fed rivers.
	Extreme sea level and coastal impacts (Sections 3.5.3, 3.5.4, and 3.5.5)	*Likely* increase in extreme high water worldwide related to trends in mean sea level in the late 20th century.	*Likely* anthropogenic influence via mean sea level contributions.	*Very likely* that mean sea level rise will contribute to upward trends in extreme sea levels. *High confidence* that locations currently experiencing coastal erosion and inundation will continue to do so due to increasing sea level, all other factors being equal. Due to *very likely* sea ice retreat, and permafrost degradation, the frequency and magnitude of the rate of Arctic coastal erosion is *likely* to increase
	Other impacts (Sections 3.5.6, 3.5.7, and 3.5.8)	*Low confidence* of global trends in large landslides in some regions. *Likely* increased thawing of permafrost with *likely* resultant physical impacts.	*Likely* anthropogenic influence on thawing of permafrost. *Low confidence* of other anthropogenic influences because of insufficient evidence for trends in other physical impacts in cold regions.	*Medium confidence* of increase in number of shallow landslides and debris flows from recently deglaciated terrain, and that high-mountain debris flows will begin earlier in the year. *Low confidence* in projected changes in the magnitude and frequency of shallow landslides in temperate and tropical regions. *Likely* that continued permafrost degradation will further decrease the stability of rock slopes, though there is *low confidence* regarding future locations and times of large rock avalanches. *Low confidence* in projected future changes in dust activity.

Table 3.2: Regional observed changes in temperature and precipitation extremes, including dryness. See Figure 3.2 for definitions of regions.

Region and Sub-region		**Tmax [WD: warm days CD: cold days]** (using 1961-1990 extreme values as reference, e.g., 90th/10th percentile)	**Tmin [WN: warm nights CN: cold nights]** (using 1961-1990 extreme values as reference, e.g., 90th/10th percentile)	**Heat waves (HW)/ Warm spells [HWDmean/max: mean/max heat wave duration WSDI: Warm spell duration index]** (using 1961-1990 extreme values as reference)	**Heavy Precipitation (HP)**	**Dryness [CDD: consecutive dry days SM: (simulated) soil moisture PDSI: Palmer-drought severity index]**
North America	All North America	*High confidence: Likely* overall increase in HD, decrease in CD (Alexander et al., 2006).	*High confidence: Likely* overall decrease in CN, increase in WN (Alexander et al., 2006).	*Medium confidence*: Increase since 1960 (Kunkel et al., 2008).	*High confidence: Likely* increase in many areas since 1950 (Trenberth et al., 2007; Kunkel et al., 2008).	*Medium confidence:* Overall slight decrease in dryness (SM, PDSI, CDD) since 1950; regional variability and 1930s drought dominate the signal (Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
	W. North America	*High confidence: Very likely* large increases in HD, large decreases in CD (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008).	*High confidence: Very likely* large decreases in unusually CN, large increases in WN (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008).	*Medium confidence*: Increase in WSDI (Alexander et al., 2006).	*Medium confidence: Spatially varying trends.* General increase, decrease in some areas, (Alexander et al., 2006).	*Medium confidence:* No overall or slight decrease in dryness (SM, PDSI, CDD) since 1950; large variability, large drought of 1930s dominates (Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
	Central North America (CNA)	*Medium confidence: Spatially varying trends.* Small increases in HD, decreases in CD in north CNA. Small decreases in HD, increases in CD in south CNA (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008).	*Medium confidence: Spatially varying trends.* Small decreases in CN, increases in WN in north CNA. Small increases in CN, decreases in WN in south CNA (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008).	*Medium confidence*: *spatially varying trends.* Some areas increase, others decrease (Alexander et al., 2006).	*High confidence: Very likely* increase since 1950 (Alexander et al., 2006).	*Medium confidence*: Decrease in dryness (SM, PDSI, CDD) and increase in mean precipitation since 1950; large variability, large drought of 1930s dominates (Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
	E. North America	*Medium confidence: Spatially varying trends.* Overall increases in WD, decreases in CD; opposite or insignificant signal in a few areas (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008).	*Medium confidence: Spatially varying trends.* Overall small decreases in CN, overall small increases in WN in NE North America. Overall small increases in CN, overall small decreases in WN in SE North America (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008).	*Medium confidence: Spatially varying trends.* Many areas increase, some areas decrease (Alexander et al., 2006).	*High confidence: Very likely* increase since 1950 (Alexander et al., 2006).	*Medium confidence*: Slight decrease in dryness (SM, PDSI, CDD) since 1950, large variability, large drought of 1930s dominates (Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
	Alaska/NW Canada	*High confidence: Very likely* large increases in WD, large decreases in CD (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008).	*High confidence: Very likely* large decreases in CN, large increases in WN (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008).	*Low confidence*: *Insufficient evidence*	*Medium confidence:* Suggestion of increase, no significant trend (Kunkel et al., 2008).	*Medium confidence:* Inconsistent trends; increase in dryness (SM, PDSI, CDD) since 1950 in part of the region. (Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).

	E. Canada, Greenland, Iceland	*High confidence: Likely* increases in WD in some areas, decrease in others. Decreases in CD in some areas, increase in others (Robeson, 2004; Alexander et al., 2006; Vincent and Mekis, 2006; Trenberth et al., 2007; Kunkel et al., 2008; Peterson et al., 2008).	*Medium confidence:* Small increases in unusually cold nights, decreases in WN in northeastern Canada. Small decreases in CN, increases in WN in southeastern and south central Canada. (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008).	*Medium confidence:* Some areas increase, most others decrease (Alexander et al., 2006).	*Medium confidence:* Increase in a few areas (Alexander et al., 2006).	*Low confidence: Insufficient evidence*
Europe	All Europe	*High confidence*: Overall *likely* increase in WD and *likely* decrease of CD over most of the continent since 1950. Strong increasing tendency in WD in most regions since 1976 onward; small or insignificant decrease in CD over same period (Alexander et al., 2006; Brown et al., 2008; see also entries for individual subregions).	*High confidence*: Overall *likely* increase in WN and *likely* decrease of CN over most of the continent since 1950. Strong increasing tendency in WN in most regions since 1976 onward; small or insignificant decrease in CN over same period (Klein Tank and Können, 2003; Alexander et al., 2006; Brown et al., 2008; see also entries for individual subregions).	*Medium confidence*: Increase of HW since 1950. Overall consistent positive trend of WSDI across Europe, but no single region with significant trends (Alexander et al., 2006). Availability of a few single studies for specific regions (see below).	*Medium confidence*: Increase in part of the region, mostly in winter, insignificant or inconsistent changes elsewhere, in particular in summer. Some inconsistencies on overall patterns between studies depending on considered indices. Most consistent signal over Central-Western Europe and European Russia (Klein Tank and Können, 2003; Haylock and Goodess, 2004; Alexander et al., 2006; Zolina et al., 2009).	*Medium confidence:* Inconsistent trends. Increase in dryness (SM, PDSI, CDD) in part of the region; insignificant, inconsistent, or no changes elsewhere. Most consistent signal for increase in dryness in Central and Southern Europe since the 1950s. No signal in Northern Europe. (Kiktev et al., 2003; Haylock and Goodess, 2004; Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
	N. Europe	*Medium confidence*: Increase in WD and decrease in CD. Consistent signals for whole region, but generally not significant at the local scale (Alexander et al., 2006). Significant increase in location parameter of yearly warmest days and insignificant trends in location parameter of yearly coldest days (Brown et al., 2008).	*Medium confidence*: Increase in WN and decrease in CN. Consistent signals over whole region but generally not significant at the local scale (Klein Tank and Können, 2003; Alexander et al., 2006; Brown et al., 2008).	*Medium confidence:* Increase of HW. Consistent tendency for increase of WSDI, but no significant trends (Alexander et al., 2006).	*Medium confidence*: Increase in winter in some areas, but often insignificant or inconsistent trends at sub-regional scale, in particular in summer (Fowler and Kilsby, 2003; Kiktev et al., 2003; Klein Tank and Können, 2003; Alexander et al., 2006; Maraun et al., 2008; Zolina et al., 2009).	*Medium confidence: Spatially varying trends.* Overall only slight or no increase in dryness (SM, PDSI, CDD), slight decrease in dryness in part of the region (Kiktev et al., 2003; Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
	Central Europe	*High confidence: Likely* overall increase in WD and *likely* decrease in CD since 1950 in most regions. Some regional and temporal variations in significance of trends. *High confidence*: *Very likely* increase in WD since 1950, 1901 and 1880 and *likely* decrease in CD since 1950 and 1901 in West-Central Europe (Alexander et al., 2006; Della-Marta et al., 2007; Laurent and Parey, 2007; Brown et al., 2008). *Medium confidence*: Lower confidence in trends in East-Central Europe compared to West-Central Europe due to lack of literature, partial lack of access to observations, overall weaker signals, and change point in trends at the end of the 1970s / beginning of 1980s. Overall tendency for increase in WD and decrease in CD, strongest increase in WD since 1976. (Alexander et al., 2006; Bartholy and Pongracz, 2007; Brown et al., 2008; Hirschi et al., 2011).	*High confidence*: *Likely* overall increase in WN and *likely* overall decrease in CN at the yearly time scale. Some regional and seasonal variations in signficance and in a few cases also the sign of the trends. *High confidence*: *Very likely* increase in WN and *very likely* decrease in CN since 1950 and 1901 in West-Central Europe (Kiktev et al., 2003; Alexander et al., 2006). *Medium confidence*: Lower confidence in trends in East-Central Europe compared to West-Central Europe due to lack of literature, partial lack of access to observations, overall weaker signals, and change point in trends at the end of the 1970s / beginning of 1980s. Overall tendency for increase in WN and decrease in CN on the yearly time scale. (Klein Tank and Können, 2003; Alexander et al., 2006; Bartholy and Pongracz, 2007; Brown et al., 2008).	*Medium confidence:* Increase in heat waves. Consistent tendency for increase of HW (WSDI) but no significant trends (Alexander et al., 2006). Significant increase in max heatwave duration HWDmax since 1880 in West-Central Europe in summer (JJA) (Della-Marta et al., 2007). Less significant signal in heatwave indices in Eastern Europe due to presence of change point (Bartholy and Pongracz, 2007; Hirschi et al., 2011).	*Medium confidence*: Increase in part of the domain, in particular in Central Western Europe and European Russia, especially in winter. Insignificant or inconsistent trends elsewhere, in particular in summer (Kiktev et al., 2003; Klein Tank and Können, 2003; Schmidli and Frei, 2005; Alexander et al., 2006; Bartholy and Pongracz, 2007; Kysely, 2009; Tomassini and Jacob, 2009; Zolina et al., 2009).	*Medium confidence: Spatially varying trends*. Increase in dryness (SM, PDSI, CDD) in part of the region but some regional variation of dryness trends and dependence of of trends on considered studies (index, time period) (Kiktev et al., 2003; Alexander et al., 2006; Bartholy and Pongracz, 2007; Sheffield and Wood, 2008a; Brazdil et al., 2009; Dai, 2011).

	S. Europe and Mediterranean	*High confidence: Likely* increase in WD and *likely* decrease in CD in most of the region. Some regional and temporal variations in significance of trends. *Likely* strongest and most significant trends in the Iberian Peninsula and Southern France (Alexander et al., 2006; Brunet et al., 2007; Della-Marta et al., 2007; Bartolini et al., 2008; Brown et al., 2008; Toreti and Desiato, 2008; Kuglitsch et al., 2010; Rodríguez-Puebla et al., 2010; Hirschi et al., 2011). *Medium confidence:* Smaller or less significant trends in Southeastern Europe and Italy due to change point in trends at the end of the 1970s / beginning of 1980s; sometimes linked with changes in sign of trends; strongest WD increase since 1976 (Bartholy and Pongracz, 2007; Bartolini et al., 2008; Toreti and Desiato, 2008; Kuglitsch et al., 2010; Hirschi et al., 2011).	*High confidence*: *Likely* increase in WN and *likely* decrease in CN in most of the region. Some regional variations in significance of trends. *Very likely* overall increase in WN and *very likely* overall decrease in CN in southwestern Europe and western Mediterranean; *likely* strongest signals in Spain and Southern France (Kiktev et al., 2003; Klein Tank and Können, 2003; Alexander et al., 2006; Brunet et al., 2007; Brown et al., 2008; Toreti and Desiato, 2008; Rodríguez-Puebla et al., 2010). *Likely* overall tendency for increase in WN and *likely* overall tendency for decrease in CN in SE Europe and eastern Mediterranean. (Kiktev et al., 2003; Klein Tank and Können, 2003; Alexander et al., 2006; Brown et al., 2008).	*High confidence: Likely* overall increase in HW in summer (JJA). Significant increase in HWDmax since 1880 in Iberian Peninsula and West-C. Europe in JJA (Della-Marta et al., 2007). Significant increase of HWDmax in Tuscany (Italy) (Bartolini et al., 2008). Significant increase in HW indices in Turkey and to a smaller extent in SE Europe and Turkey in JJA (Kuglitsch et al., 2010). Less significant signal in HW indices in SE Europe due to presence of change point in trends (Bartholy and Pongracz, 2007; Hirschi et al., 2011).	*Low confidence*: Inconsistent trends within domain and across studies. (Kiktev et al., 2003; Klein Tank and Können, 2003; Alexander et al., 2006; García et al., 2007; Pavan et al., 2008; Zolina et al., 2009; Rodrigo, 2010).	*Medium confidence:* Overall increase in dryness (SM, PDSI, CDD), but partial dependence on index and time period (Kiktev et al., 2003; Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
Africa	All Africa	*Low confidence: Insufficient evidence* (*lack of literature* in many regions)	*Low confidence: Insufficient evidence* (*lack of literature* in many regions)	*Low confidence: Insufficient evidence* (*lack of literature* for the overall continent)	*Low confidence:* Partial lack of data and literature and inconsistent patterns in existing studies (New et al., 2006; Aguilar et al., 2009; Camberlin et al., 2009).	*Medium confidence:* Overall increase in dryness (SM, PDSI); regional variability, 1970s prolonged Sahel drought dominates. (Sheffield and Wood, 2008a; Dai, 2011). No apparent continent-wide trends of change in rainfall over the 20th century, although there was a continent wide drought in 1983 and 1984 (Hulme et al., 2001). Wet season arrives 9–21 days later, large inter-annual variability of wet season start, local scale geographical variability (Kniveton et al., 2009).
	W. Africa	*Medium confidence:* Significant increase in temperature of warmest day and coldest day, significant increase in frequency of warm days, and significant decrease in frequency of cold days in western central Africa and Guinea Conakry (Aguilar et al., 2009); *lack of literature* for other parts of the region.	*Medium confidence:* Decreases in frequency of CN in western central Africa (significant) and Guinea Conakry (not significant) (Aguilar et al., 2009); *lack for literature* for other parts of the region. *High confidence: Likely* increases in WN (Trenberth et al., 2007; Aguilar et al., 2009).	*Low confidence: Insufficient evidence* (*lack of literature* for the overall continent).	*Medium confidence:* Precipitation from heavy events has decreased (Western Central Africa, Guinea Conakry) but low spatial coherence (Aguilar et al., 2009), rainfall intensity increased (New et al., 2006).	*Medium confidence:* 1970s prolonged Sahel drought dominates, conditions are still drier (SM, PDSI, precipitation anomalies) than during the humid 1950s (L'Hôte et al., 2002; Dai et al., 2004; Sheffield and Wood, 2008a; Dai, 2011). Recent years characterized by a greater interannual variability than previous 40 years, western Sahel remaining dry and the eastern Sahel returning to wetter conditions (Ali and Lebel, 2009), dry spell duration increased (New et al., 2006).

	E. Africa	*Low confidence*: *Lack of literature* for most of the domain; *Spatially non-uniform trends* in daytime temperature, some areas with cooling (King'uyu et al., 2000). *Medium confidence:* Increases in WD, decreases in CD in southern tip of domain (Trenberth et al., 2007).	*Medium confidence:* Spatially non-uniform, rise of night-time temperature at several locations, but with many coastal areas and stations near large water bodies showing a significant decrease (King'uyu et al., 2000); Decreases in CN, increases in WN, in southern tip of the domain (Trenberth et al., 2007).	*Low confidence: Insufficient evidence* (*lack of literature* for the overall continent)	*Medium confidence:* Decrease (Trenberth et al., 2007). Inter-annual variations in the onset date of the rainy season have had the biggest impact on seasonal rains (Camberlin et al., 2009).	*Low confidence: Spatially varying trends* in dryness (SM, PDSI) (Sheffield and Wood, 2008a; Dai, 2011).
	S. Africa	*Medium confidence:* Increases in WD, decreases in CD (Trenberth et al., 2007).	*Medium confidence:* Decreases in CN, increases in WN (Alexander et al., 2006). Regional variations, increase at several locations, but with many coastal areas and stations near large water bodies showing a significant decrease (King'uyu et al., 2000).	*Low confidence: Insufficient evidence* (*lack of literature* for the overall continent)	*Medium confidence:* Rainfall intensity increased (Kruger, 2006; New et al., 2006; Trenberth et al., 2007).	*Medium confidence:* Slight dry spell duration increase (Alexander et al., 2006; Kruger, 2006; New et al., 2006) General increase in dryness (SM, PDSI) (Sheffield and Wood, 2008a; Dai, 2011).
	Sahara	*Low confidence: Lack of literature.*	*Medium Confidence:* Increases in unusually WN (Trenberth et al., 2007). *Lack of literature* on trends in unusually CN.	*Low confidence: Insufficient evidence* (*lack of literature* for the overall continent)	*Low confidence: Insufficient evidence*	*Low confidence:* Limited data, spatial variation of the trends (Dai, 2011).
Central and South America	All SA	*Low confidence: Insufficient evidence*	*Low confidence: Insufficient evidence*	*Low confidence: Insufficient evidence*	*Low confidence: Insufficient evidence*	*Low confidence: Spatially varying trends, Inconsistencies between studies* (Sheffield and Wood, 2008a; Dai, 2011).
	C. America & Northern SA	*Medium confidence:* Increases in WD, decreases in CD (Aguilar et al., 2005; Alexander et al., 2006; Brown et al., 2008).	*Medium confidence:* Decreases in CN, increases in WN (Aguilar et al., 2005; Alexander et al., 2006; Brown et al., 2008).	*Low confidence: Spatially varying trends.* A few areas increase, a few others decrease (Aguilar et al., 2005; Alexander et al., 2006).	*Medium confidence: Spatially varying trends.* Increase in many areas, decrease in a few areas, (Aguilar et al., 2005; Alexander et al., 2006).	*Low confidence: Spatially varying trends, Inconsistencies in trends* in dryness (SM, PDSI, CDD). (Aguilar et al., 2005; Sheffield and Wood, 2008a; Dai, 2011).
	Amazon	*Low confidence: Insufficient evidence.*	*Low confidence: Insufficient evidence* (Alexander et al., 2006; Dufek et al., 2008).	*Low confidence:* Insufficient evidence	*Medium confidence: Spatially varying trends.* Increase in many areas, decrease in a few areas (Alexander et al., 2006; Haylock et al., 2006).	*Low confidence: Spatially varying trends, Mixed results.* Slight decrease of CDD (Dufek et al., 2008). Tendency to decreased dryness in much of region, but some opposite trends and inconsistencies between studies (Sheffield and Wood, 2008a; Dai, 2011).
	Northeastern Brazil	*Medium confidence:* Increases in WD (Silva and Azevedo, 2008).	*Medium confidence:* Increases in WN (Silva and Azevedo, 2008).	*Low confidence: Insufficient evidence*	*Medium confidence:* Increase in many areas, decrease in a few areas, (Alexander et al., 2006; Haylock et al., 2006; Santos and Brito, 2007; Silva and Azevedo, 2008; Santos et al., 2009).	*Low confidence: Spatially varying trends.* Inconsistent trends in CDD (Santos and Brito, 2007; Dufek et al., 2008; Silva and Azevedo, 2008; Santos et al., 2009).Inconsistent trends in dryness (SM, PDSI) between studies (Sheffield and Wood, 2008a; Dai, 2011).

	SE South America	*Medium confidence: Spatially varying trends.* Increases in WD in some areas, decrease in others. Decreases in CD in some areas, increase in others, (Rusticucci and Barrucand, 2004; Vincent et al., 2005; Alexander et al., 2006; Brown et al., 2008; Rusticucci and Renom, 2008; Marengo et al., 2009b).	*Medium confidence:* Decreases in CN, increases in WN (Rusticucci and Barrucand, 2004; Vincent et al., 2005; Alexander et al., 2006; Brown et al., 2008; Rusticucci and Renom, 2008; Marengo et al., 2009b).	*Low confidence: Spatially varying trends.* Some areas increase, others decrease (Alexander et al., 2006).	*Medium Confidence:* Increase (Alexander et al., 2006; Dufek et al., 2008; Sugahara et al., 2009; Penalba and Robeldo, 2010).	*Low Confidence:* Slight increase in dryness, large variability (Haylock et al., 2006; Dufek and Ambrizzi, 2008; Dufek et al., 2008; Penalba and Robeldo, 2010; Llano and Penalba, 2011). Decrease in dryness (SM, PDSI) in much of region (Sheffield and Wood, 2008a; Dai, 2011).
	W. Coast SA	*Medium confidence:* Increases in WD in some areas, decrease in others. Decreases in CD in some areas, increase in others, (Rosenbluth et al., 1997; Vincent et al., 2005; Alexander et al., 2006).	*Medium confidence:* Decreases in CN, increases in WN (Rosenbluth et al., 1997; Vincent et al., 2005; Alexander et al., 2006).	*Low confidence: Insufficient evidence*	*Medium confidence: Spatially varying trends.* Decrease in many areas, increase in a few areas (Alexander et al., 2006; Haylock et al., 2006).	*Medium Confidence:* Slight increased dryness (SM, PDSI, CDD) in most areas (Dufek et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
Asia	All Asia	*Medium confidence:* Increase in WD, decrease in CD	*Medium confidence:* Increase in WN, decrease in CN	*Low confidence: Insufficient evidence*	*Low confidence: Insufficient evidence*	*Low confidence: Spatially varying trends*. Some areas present consistent increases, respectively decreases, in dryness (SM, PDSI, CDD) (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011) .
Asia	N. Asia	*High confidence: Likely* increases in WD, decreases in CD (Alexander et al., 2006).	*Medium confidence:* Decreases in CN, increases in WN (Alexander et al., 2006; Trenberth et al., 2007).	*Medium confidence: Spatially varying trends.* General increase, decrease in a few areas (Alexander et al., 2006).	*Medium confidence:* Increase (Trenberth et al., 2007). Some increase in western Russia, especially in winter (DJF), 1950-2000 (Zolina et al., 2009).	*Low confidence: Spatially varying trends.* Tendency for increased dryness (SM, PDSI, CDD) in central and northeastern N. Asia, other areas decreased dryness (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
Asia	Central Asia	*High confidence: Likely* increases in WD, decreases in CD (Alexander et al., 2006; Trenberth et al., 2007).	*High confidence:* Decreases in CN, increases in WN (Alexander et al., 2006; Trenberth et al., 2007).	*Medium confidence:* Increase in a few areas, insufficient evidence elsewhere (Alexander et al., 2006).	*Low confidence: Spatially varying trends.* Increase in a few areas, decrease in a few areas. (Alexander et al., 2006).	*Low confidence: Spatially varying trends* in dryness (SM, PDSI, CDD); partial lack of coverage in some studies (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
Asia	East Asia	*High confidence: Likely* increases in WD, decreases in CD (Trenberth et al., 2007; Ding et al., 2010).	*Medium confidence:* Decreases in CN, increases in WN (Alexander et al., 2006).	*Medium confidence:* Increase in warm season heat waves in China (Ding et al., 2010), but decline in all warm spells (Alexander et al., 2006).	*Low confidence: Spatially varying trends.* Increase in a few areas, decrease in a few areas (Alexander et al., 2006).	*Medium confidence:* Overall tendency for increased dryness (SM, PDSI, CDD); few areas with opposite trends (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
Asia	S.E. Asia	*Medium confidence:* Increases in WD, decreases in CD, northern part (Alexander et al., 2006).	*Medium confidence:* Decreases in CN, increases in WN, northern part. (Alexander et al., 2006).	*Low confidence: Insufficient evidence*	*Low confidence: Spatially varying trends.* Some areas increase, some areas decrease (Alexander et al., 2006).	*Low confidence: Spatially varying trends, Inconsistent trends* in dryness (SM, PDSI) between studies (Sheffield and Wood, 2008a; Dai, 2011).

	S. Central Asia	*Medium confidence:* Increase in WD and decrease in CD (Alexander et al., 2006).	*Medium confidence:* Decreases in CN, increases in WN (Alexander et al., 2006).	*Low confidence*: *Insufficient evidence*	*Low Confidence:* Mixed signal in India (Alexander et al., 2006).	*Low confidence: Inconsistent signal for different studies and indices.* Decrease in CDD over India (Alexander et al., 2006). Increased dryness (SM, PDSI) in central India (Sheffield and Wood, 2008a; Dai, 2011).
	Tibetan Plateau	*High confidence: Likely* increase in WD and decrease in CD (Alexander et al., 2006).	*High confidence: Likely* decreases in CN, increases in WN (Alexander et al., 2006).	*Low confidence:* Increase in a few areas, insufficient evidence elsewhere (Alexander et al., 2006).	*Low confidence: Insufficient evidence.*	*Low confidence: Lack of studies.* Tendency to decreased dryness (PDSI, SM) in Dai (2011).
	W. Asia	*High confidence: More likely than not* decrease in CD and *very likely* increase in WD (Choi et al., 2009; Rahimzadeh et al., 2009; Rehman, 2010).	*High confidence: Likely* decrease in CN and likely increase in WN (Choi et al., 2009; Rehman, 2010).	*Medium confidence:* Increase in warm spells (Alexander et al., 2006).	*Medium confidence:* decrease in heavy precipitation events. (Kwarteng et al., 2009; Rahimzadeh et al., 2009).	*Low confidence:* Mixed results. (Kwarteng et al., 2009; Rahimzadeh et al., 2009).
Australia/New Zealand	N. Australia	*High confidence: Very likely* increases in WD, decreases in CD. Weaker trends in northwest (Alexander et al., 2006; Trenberth et al., 2007).	*High confidence: Very likely* decreases in CN, increases in WN (Alexander et al., 2006).	*Low confidence:* Insufficient studies for assessment	*Low confidence:* Insufficient studies for assessment.	*Medium confidence:* Decrease in dryness (SM, PDSI) in northwest since mid-20th century (Sheffield and Wood, 2008a; Dai, 2011).
	S. Australia/NZ	*High confidence: Very likely* increases in WD, decreases in CD (Alexander et al., 2006; Trenberth et al., 2007). NZ positive trends vary across country, related to circulation changes (Chambers and Griffiths, 2008; Mullan et al., 2008).	*High confidence: Very likely* decreases in CN, increases in WN (Alexander et al., 2006). General decrease in frosts in NZ but trends vary across country, related to circulation changes (Chambers and Griffiths, 2008; Mullan et al., 2008).	*Medium confidence:* increase in warm spells across southern Australia (Alexander and Arblaster, 2009).	*High confidence: Likely* increase in heavy precipitation in many areas, except where mean precipitation has decreased (CSIRO and Bureau of Meteorology, 2007; Gallant et al., 2007; Alexander and Arblaster, 2009). NZ trends are positive in western N. and S. Islands and negative in east of country, and are strongly correlated with changes in mean rainfall (Mullan et al., 2008).	*Medium confidence:* Increase in dryness (SM, PDSI, CDD) in southeastern part and southwestern tip of continent since mid-20th century; decrease in dryness in central part of continent (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).

Table 3.3: Projected regional changes in temperature and precipitation (including dryness) extremes. See Figure 3.2 for definitions of regions. Projections are for the end of the 21st century vs end of the 20th century (i.e., 1961-1990 or 1980-1999 vs 2071-2100 or 2080-2099) and for the A2/A1B emissions scenario. Codes for the source of modelling evidence: G: Based on GCM simulations. **G**: multi-GCM. R: Based on RCM simulations. **R**: multi-GCM. R: multi-RCM. T06 stands for Tebaldi et al. (2006), SW08 stands for Sheffield and Wood (2008b), and OS11 stands for Orlowsky and Seneviratne (2011).

Region and Sub-Region		**Tmax [WD: warm days CD: cold days] (using 1961-1990 extreme values as reference, e.g., 90th/10th percentile)**		**Tmin [WN: warm nights CN: cold nights] (using 1961-1990 extreme values as reference, e.g., 90th/10th percentile)**		**Heat waves (HW) / Warm spells [HWDmean/max: mean/max heat wave duration WSDI: Warm spell duration index] (using 1961-1990 extreme values as reference)**		**Heavy precipitation (HP) [HPD: heavy precipitation days, e.g., precipitation > 95th percentile %DP10: percentage of days with precipitation > 10mm HPC: heavy precipitation contribution e.g., fraction from precipitation > 95th percentile] (using 1961-1990 extreme values as reference except in case of %DP10)**		**Dryness [CDD: consecutive dry days SM: (simulated) soil moisture PDSI: Palmer-drought severity index]**	
North America	All North America	*High confidence*: HD *very likely* to increase & CD *very likely* to decrease in all regions (Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; OS11). *Medium confidence*: Largest increases of HD in summer and fall particularly over the US; largest decrease of CD in Canada in fall and winter (OS11).	**G**	*High confidence*: WN *very likely* to increase & CN *very likely* to decrease over all regions (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; OS11). *Medium confidence :* Largest increase in WN and decrease in CN in summer, particularly in the US (OS11).	**G**	*High confidence: Likely* more frequent and longer heat waves & warm spells over all regions (T06; Christensen et al., 2007; Karl et al., 2008; OS11).	**G**	*Low to high confidence: Likely* increase in HPD and HPC over Canada and Alaska; smaller and less consistent changes in south, but inconsistencies between %DP10 (decreases in winter spring) and other indices (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; OS11).	**G**	*Low to medium confidence. Medium confidence* regarding increase in CDD, and SM drought in Texas and N. Mexico (T06; SW08; OS11). *Low confidence*: Inconsistent change in other regions (SM, CDD) (T06; SW08; OS11).	**G**
	W. North America	*High confidence*: HD *very likely* to increase & CD *very likely* to decrease in all seasons (Christensen et al., 2007; Karl et al., 2008; Clark et al., 2011; OS11). *Medium confidence*: Overall weaker signal in spring and winter for both CD and HD (OS11).	**G**	*High confidence*: WN *very likely* to increase & CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; OS11). *Medium confidence:* Largest WN increases and CN decreases in summer (OS11).	**G**	*High confidence: Likely* more frequent and longer heat waves & warm spells (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; Clark et al., 2011; OS11).	**G**	*Low to medium confidence:* Increase of HPD/HPC over northern part of domain (Canada); no signal or inconsistent signal over southern part of domain (T06; OS11).	**G**	*Low confidence:* inconsistent signal in CDD and SM changes (T06; SW08; OS11).	**G**
	Central North America	*High confidence*: HD *very likely* to increase & CD *very likely* to decrease in all seasons (Christensen et al., 2007; Karl et al., 2008; Clark et al., 2011; OS11). *Medium confidence*: Weaker signal for CD in spring and winter (OS11).	**G**	*High confidence*: WN *very likely* to increase & CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; OS11).	**G**	*High confidence: Likely* more frequent and longer heat waves & warm spells. (T06; Christensen et al., 2007; Karl et al., 2008; Clark et al., 2011; OS11).	**G**	*Low confidence*: Inconsistent or no signal (T06; OS11).	**G**	*Medium confidence:* Increase in CDD and decrease of SM in southern part of the domain (SW08; OS11); Low confidence: inconsistent signal elsewhere (OS11).	**G**
	E. North America	*High confidence*: HD *very likely* to increase & CD *very likely* to decrease in all seasons (Christensen et al., 2007; Karl et al., 2008; Clark et al., 2011; OS11). *Medium confidence*: Largest HD increase in summer and fall; weaker CD decrease in spring (OS11).	**G**	*High confidence*: WN *very likely* to increase & CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; OS11). *Medium confidence:* Largest WN increases and CN decreases in summer (OS11).	**G**	*High confidence: Likely* more frequent and longer heat waves & warm spells (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; Clark et al., 2011; OS11).	**G**	*Medium confidence*: Increase of HPD/HPC in northern part of domain but no signal or inconsistent signal in southern part (T06; OS11).	**G**	*Low confidence:* inconsistent signal in CDD, some consistent decrease of SM (SW08; OS11).	**G**

	Alaska / NW Canada	*High confidence*: HD *very likely* to increase & CD *very likely* to decrease (Christensen et al., 2007; Karl et al., 2008; OS11). *Medium confidence*: strongest increase of HD in the fall (OS11).	**G**	*High confidence*: WN *very likely* to increase & CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; OS11).	**G**	*High confidence: Likely* more frequent and longer heat waves & warm spells (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; Karl et al., 2008; OS11).	**G**	*High confidence*: *Likely* increase of HPD and HPC (T06; Kharin et al., 2007; OS11).	**G**	*Low confidence:* inconsistent signal in change of CDD and SM (T06; SW08; OS11).	**G**
	E. Canada, Greenland, Iceland	*High confidence*: HD *very likely* to increase & CD *very likely* to decrease (Christensen et al., 2007; Karl et al., 2008; OS11). *Medium confidence*: strongest increase of HD in fall and winter (in summer in Greenland), weakest in spring, weaker increase of CD in summer (OS11).	**G**	*High confidence*: WN *very likely* to increase & CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007; OS11).	**G**	*High confidence: Likely* more frequent and longer heat waves & warm spells (T06; OS11).	**G**	*High confidence*: *Likely* increase of HPD and HPC (T06; Kharin et al., 2007; OS11).	**G**	*Low confidence:* inconsistent signal in CDD and/or SM changes (T06; SW08; OS11).	**G**
Europe	All Europe	*High confidence*: HD *very likely* to increase – largest increases in summer and C/S Europe & smallest in N Europe (Scandinavia) (Goubanova and Li, 2007; Kharin et al., 2007; Kjellstrom et al., 2007; OS11; Koffi and Koffi, 2008) and cold days *very likely* to decrease (OS11). *Medium confidence:* Changes in higher quantiles of Tmax generally greater than changes in lower quantiles of Tmax in summer in Central Europe and Mediterranean (Diffenbaugh et al., 2007; Kjellstrom et al., 2007; Fischer and Schär, 2009; Fischer and Schär, 2010; OS11).	**G** **<u>R</u>**	*High confidence*: CN *very likely* to decrease – largest decreases in winter & E Europe & Scandinavia (Goubanova and Li, 2007; Kjellstrom et al., 2007; Sillmann and Roeckner, 2008).WN *very likely* to increase (T06; Kharin et al., 2007; OS11).	**G** **<u>R</u>**	*High confidence:* HWD *likely* to increase (also increases in intensity & frequency) but little change over Scandinavia (Beniston et al., 2007; Christensen et al., 2007; Kysely and Beranova, 2009; Clark et al., 2011; OS11).	**G** **<u>R</u>**	*Low to high confidence, depending on regions: Very likely* increases in HPD, %DP10, and decreases in return periods of long (5-day) and short (1-day) events in N. Europe particularly in winter; lower confidence in changes in C. Europe and the Mediterranean (T06; Beniston et al., 2007; Fowler et al., 2007; Kharin et al., 2007; Sillmann and Roeckner, 2008; Kendon et al., 2010; OS11). *Likely* increase in HPC in some regions (Boberg et al., 2009; Kendon et al., 2010). *Likely* greater changes in extremes than mean in many regions.Increase in HP intensity (& increase in HPC) despite decrease in summer mean in some regions – e.g. C. Europe (Beniston et al., 2007; Fowler et al., 2007; Haugen and Iverson, 2008; May, 2008; Kysely and Beranova, 2009).	**G** **<u>R</u>**	*Medium confidence:* European area affected by stronger dryness (reduced SM and CDD) with largest and most consistent changes in Mediterranean Europe (T06; Burke and Brown, 2008; May, 2008; SW08; Sillmann and Roeckner, 2008; OS11).	**G** R

	N. Europe	*High confidence: Very Likely* increase in frequency of HD, but smaller than in Central and Southern Europe (Fischer and Schär, 2010; OS11). *Very likely* decrease in CD (OS11). *Medium confidence:* Changes in lower quantiles of Tmax generally greater than for changes in higher quantiles of Tmax in fall, winter and spring in Scandinavia and Northeastern Europe (OS11).	**G** **R**	*High confidence*: CN *very likely* to decrease (Goubanova and Li, 2007; Kjellstrom et al., 2007; Sillmann and Roeckner, 2008); WN *very likely* to increase (T06; OS11). *Medium confidence:* Changes in lower quantiles of Tmin generally greater than changes in higher quantiles of Tmin in Scandinavia and Northeastern Europe (Goubanova and Li, 2007; Kjellstrom et al., 2007; OS11).	**G** **R**	*High confidence*: HWD *likely* to increase, but summer increases smaller than in S Europe and little change over Scandinavia (Beniston et al., 2007; Kysely and Beranova, 2009; Clark et al., 2011; OS11).	**G** **R**	*High confidence: Very likely* increases in HP (intensity and frequency) and %DP10 north of 45N in winter (Frei et al., 2006; T06; Beniston et al., 2007; Kendon et al., 2008; OS11).	**G** R	*Medium confidence*: No major changes in dryness (CDD, SM deficit) in Northern Europe (T06; SW08; Sillmann and Roeckner, 2008; OS11).	**G**
	Central Europe	*High confidence: Very likely* increase in frequency and intensity of HD (Fischer and Schär, 2010; OS11) and decrease in frequency of CD (OS11). *Medium confidence:* Changes in higher quantiles of Tmax much larger than changes in lower quantiles of Tmax in summer; results in very large increase in Tmax variability (Seneviratne et al., 2006; Diffenbaugh et al., 2007; Kjellstrom et al., 2007; Fischer and Schär, 2009; Fischer and Schär, 2010; OS11).	**G** **R**	*High confidence*: CN *very likely* to decrease (Goubanova and Li, 2007; Kjellstrom et al., 2007; Sillmann and Roeckner, 2008); WN *very likely* to increase (T06; OS11).	**G** R	*High confidence*: HWD *likely* to increase (Beniston et al., 2007; Kysely and Beranova, 2009; Fischer and Schär, 2010; OS11).	**G** **R**	*High confidence: Likely* increases in HP (intensity and frequency) in large part of the region in winter (Frei et al., 2006; Beniston et al., 2007; Kendon et al., 2008). *Medium confidence:* Inconsistent evidence for summer: increase in HP in summer evident in RCMs (Christensen and Christensen, 2003; Frei et al., 2006) vs no signal in GCMs (OS11).	**G** **R**	*Medium confidence:* Increase in dryness (CDD, SM) in Central Europe (Seneviratne et al., 2006; T06; OS11). *Medium confidence:* Increase in short-term droughts (SW08).	**G** R
	S. Europe and Mediterranean	*High confidence: Very likely* increase in frequency and intensity of HD (Fischer and Schär, 2009; Giannakopoulos et al., 2009; Fischer and Schär, 2010; OS11) and decrease in frequency of cold days (OS11). *High confidence*: Number of days with combined hot summer days (>35°C) and tropical nights (>20°C) *very likely* to increase (Fischer and Schär, 2010). *Medium confidence:* Changes in higher quantiles of Tmax greater than changes in lower quantiles of Tmax in summer (Diffenbaugh et al., 2007; Kjellstrom et al., 2007; Fischer and Schär, 2009; Fischer and Schär, 2010; OS11).	**G** **R**	*High confidence*: CN *very likely* to decrease (Goubanova and Li, 2007; Kjellstrom et al., 2007; Sillmann and Roeckner, 2008). *High confidence*: WN *very likely* to increase (T06; Sillmann and Roeckner, 2008; Giannakopoulos et al., 2009; OS11). *High confidence*: Tropical nights *very likely* to increase (Sillmann and Roeckner, 2008). Number of days with combined hot summer days (>35°C) and tropical nights (>20°C) *very likely* to increase (Fischer and Schär, 2010). *Medium confidence*: Changes in higher quantiles of Tmin generally greater than changes in lower quantiles of Tmin in summer in the Mediterranean (Diffenbaugh et al., 2007; OS11).	**G** **R**	*High confidence: Likely* increase in HWD (also increases in intensity and frequency) - *likely* largest increases in SW, S & E (Beniston et al., 2007; Diffenbaugh et al., 2007; Koffi and Koffi, 2008; Giannakopoulos et al., 2009; Fischer and Schär, 2010; OS11; Clark et al., 2011).	**G** **R**	*Low confidence*: Inconsistent change in HP intensity and %D10, depends on region and season; increase in HP intensity in all seasons except summer over parts of the region, but decrease in other parts, e.g., Iberian Peninsula (T06; Goubanova and Li, 2007; Giorgi and Lionello, 2008; Giannakopoulos et al., 2009).	**G** **R**	*Medium confidence:* Increase in dryness (CDD, SM) in Mediterranean (T06; Beniston et al., 2007; SW08; Giannakopoulos et al., 2009; OS11).	**G** **R**

Africa	All Africa	*High confidence*: HD *likely* to increase & CD *likely* to decrease in all regions (Kharin et al., 2007; OS11). *Medium confidence:* Increase in HD largest in summer and fall (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; Kharin et al., 2007; OS11) and CN *likely* to decrease (OS11).	**G**	*High confidence: Likely* increase of HWD but weaker seasonal signal and ~0-30S (Clark et al., 2006; TS06; OS11).	**G**	*Low to high confidence*: Inconsistent change or no signal in HP indicators across much of continent (T06; OS11). Strongest and most consistent signal is *very likely* increase in HP in E. Africa (TS06; OS11; Shongwe et al., 2011).	**G**	*Low to medium confidence*: *Low confidence* in most regions, *medium confidence* of increase in dryness (CDD, SM) in South Africa except eastern part (T06; SW08; OS11).	**G**
	W. Africa	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11).	**G**	*High confidence: Likely* increase of HWD but weaker seasonal signal and ~0-30S (Clark et al., 2006; TS06; OS11).	**G**	*Low to medium confidence*: Slight or no change in HP indicators, inconsistent sign within region; low agreement in tropical part of region (T06; OS11).	**G**	*Low confidence*, inconsistent signal of change in CDD and SM (T06; OS11).	**G**
	E. Africa	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN likely to decrease (OS11).	**G**	*High confidence: Likely* increase of HWD but weaker seasonal signal and ~0-30S (Clark et al., 2006; TS06; OS11).	**G**	*High confidence*: *Likely* increase in HPD and HPC (TS06; OS11; Shongwe et al., 2011).	**G**	*Low confidence*, inconsistent signal of change in CDD and SM (TS06; OS11; Shongwe et al., 2011).	**G**
	S. Africa	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11).	**G**	*High confidence: Likely* increase of HWD but weaker seasonal signal (Clark et al., 2006; TS06; OS11).	**G**	*Low confidence*: Lack of agreement/signal in %DP10 and other HP indicators for the region as a whole (T06; OS11). Some evidence of increased HP intensity in the SE (Hewitson and Crane, 2006; Rocha et al., 2008; Shongwe et al., 2009).	**G**	*Medium confidence*: dryness (CDD, SM) increase, except Eastern part (T06; Shongwe et al., 2009; OS11).	**G**
	Sahara	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11).	**G**	*High confidence: Likely* increase of HWD but weaker seasonal signal (T06; OS11).	**G**	*Low confidence*: Low agreement/no signal in %DP10 and other HP indicators (T06; OS11).	**G**	*Low confidence*, inconsistent signal of change in CDD and SM (T06; OS11).	**G**
Central and South America	All South America	*High confidence*: HD *likely* to increase & CD *likely* to decrease in all regions (Kharin et al., 2007; OS11).	**G**	*High confidence*: WN *likely* to increase (T06; Kharin et al., 2007; OS11) and CN *likely* to decrease (OS11).	**G**	*Medium to high confidence: Likely* increases of HWDmax on annual time scale in most regions of South America; only *medium confidence* in HWDmax increase in Southeastern South America (T06; OS11). *Medium confidence:* Changes in HWDmean generally smaller and less consistent than HWDmax changes (OS11).	**G**	*Low to medium confidence*: Inconsistent sign of change of HP indicators across continent; some regions with model agreement (T06; OS11).	**G**	*Low to medium confidence*: inconsistent signal except for dryness increase (CDD and SM) in Northeastern Brazil and Central America (SW08; OS11).	**G**
	Central America and northern South	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11).	**G**	*High confidence: Likely* increases of HWDmax on annual time scale (T06; OS11).	**G**	*Low confidence*: *Lack of agreement* between models regarding changes in %DP10 and other HP indicators (Kamiguchi et al., 2006; T06; Campbell et al., 2011; OS11).	**G** R	*Low to medium confidence:* increase in dryness (CDD, SM) in Central America, inconsistent signal in CDD and SM in northern South America (Kamiguchi et al., 2006; T06; Campbell et al., 2011; OS11).	**G** R

	Amazon	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: *Very likely* increase of WN (T06; Uchiyama et al., 2006; Marengo et al., 2009a; OS11) and CN *likely* to decrease (OS11).	**G** R	*High confidence: Likely* increases of HWDmax on annual time scale (T06; Uchiyama et al., 2006; OS11).	**G**	*Medium confidence : Increase of precip >95th* (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; OS11).	**G** R	*Low confidence:* inconsistent signal SM (SW08; OS11) and inconsistent signal in CDD (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; OS11).	**G** R
	Northeastern Brazil	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence: Likely* increase of WN (T06; Uchiyama et al., 2006; Marengo et al., 2009a; OS11) and CN *likely* to decrease (OS11).	**G** R	*High confidence: Likely* increases of HWDmax on annual time scale (T06; Uchiyama et al., 2006; OS11).	**G**	*Low confidence: Slight or no change in %DP1 and other HP indicators* (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; OS11).	**G** R	*Medium confidence:* Dryness (CDD, SM deficit) increase (Kamiguchi et al., 2006; T06; SW08; Marengo et al., 2009a; Sorensson et al., 2010; OS11).	**G** R
	Southeastern South America	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence: Very likely* increase of WN (T06; Uchiyama et al., 2006; Marengo et al., 2009a; OS11) and CN *likely* to decrease (OS11).	**G** R	*Medium confidence*: Weak tendency for increase of HWDmax, but robust sign across majority of models (T06; Uchiyama et al., 2006; OS11).	**G**	*Medium confidence:* Increase of %DP10, precipitation intensity and precip>95th (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Nunez et al., 2009; OS11).	**G** R	*Low confidence*: inconsistent signal in SM (OS11). *Low confidence* inconsistent signal in CDD (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; OS11).	**G** R
	Western Coast of South America	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence: Likely* increase of WN (T06; Uchiyama et al., 2006; Marengo et al., 2009a; OS11) and CN *likely* to decrease (OS11).	**G** R	*High confidence: Likely* increases of HWDmax on annual time scale (T06; OS11).	**G**	*Medium confidence:* Increase in %DP10, precip>95th and other HP indicators in the tropics. *Low confidence* for the extratropics (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; OS11).	**G** R	*Low confidence:* inconsistent change of SM across domain (OS11) ,decrease in CDD in the tropics and an increase in the extratropics (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; OS11). *Medium confidence:* Increase in CDD and decrease in SM in south-west South America (SW08; OS11).	**G** R

Asia	All Asia	*High confidence*: HD *likely* to increase & CD *likely* to decrease in all regions (Kharin et al., 2007; OS11).	**G**	*High confidence*: WN *likely* to increase (T06; Kharin et al., 2007; OS11) and CN *likely* to decrease (OS11).	**G**	*Low to high confidence:* Increases of HWDmax *likely* in most regions (continent) on the annual time scale; inconsistent signal in Indonesia, Philippines, Malaysia, Papua New Guinea and neighbouring islands (T06; OS11). Weaker signals on the seasonal time scale over continent (OS11).	**G**	*Low to high confidence* depending on region: *High confidence* regarding *likely* increase of HP in N. Asia, *Medium confidence* regarding increase of HP in SE. Asia and E Asia and *Low confidence* regarding increase of HP in S and W Asia and Tibetan plateau (T06; OS11).	**G**	*Low confidence:* Inconsistent change in CDD and SM between models in large part of domain; (T06; OS11).	**G**
	N. Asia	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11).	**G**	*High confidence: Likely* increases of HWD (T06; OS11).	**G**	*High confidence*: *Likely* increase of HP (T06; OS11) including more frequent & intense HPD over most regions (Emori and Brown, 2005; Kamiguchi et al., 2006).	**G**	*Low confidence:* Inconsistent change in dryness (CDD, SM) between models in large part of domain; (T06; SW08; OS11).	**G**
	Central Asia	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11).	**G**	*High confidence: Likely* increases of HWD (T06; OS11).	**G**	*Low confidence*: Inconsistent signal in models regarding changes in HP (T06; OS11).	**G**	*Low confidence*: Inconsistent signals across indices (CDD, SM) (T06; SW08; OS11).	**G**
	East Asia	*High confidence*: HD *likely* to increase & CD *likely* to decrease across the region (Clark et al., 2011; OS11), including in Korea (Boo et al., 2006; Im and Kwon, 2007; Im et al., 2008; Koo et al., 2009; Im et al., 2011).	**G** R	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11), including in Korea (Boo et al., 2006; Koo et al., 2009; Im et al., 2011).	**G** R	*High confidence: Likely* increases of HWD (T06; Clark et al., 2011; OS11).	**G**	*Medium confidence*: Increases in HP (less consistent in %DP10 than other indicators) across the region (T06; OS11), including increase in Japan and Korea (Emori and Brown, 2005; Kimoto et al., 2005; Boo et al., 2006; Kamiguchi et al., 2006; Kusunoki and Mizuta, 2008; Kitoh et al., 2009; Su et al., 2009; Kim et al., 2010; Im et al., 2011).	**G** R	*Low confidence:* Inconsistent signal across indices (CDD, SM) (T06; SW08; OS11).	**G**
	S.E. Asia	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11).	**G**	*Low to high confidence:* Inconsistent signal in Indonesia, Philippines, Malaysia, Papua New Guinea and neighbouring islands; *likely* increases of HWDmax on the annual time scale over continental (T06; OS11).	**G**	*Medium confidence*: Inconsistent signal in change of %DP10 across models. (T06; OS11) but more frequent & intense HPD suggested by other indicators over most regions especially non-continental parts (Emori and Brown, 2005; Kamiguchi et al., 2006; OS11).	**G**	*Low confidence:* inconsistent signal of change in CDD and/or SM (T06; SW08; OS11).	**G**
	S. Asia	*High confidence*: HD *likely* to increase & CD *likely* to decrease (Kumar et al., 2006; Rajendran and Kitoh, 2008; OS11).	**G** R	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11). *Medium confidence:* Extreme nighttime temperature warms faster than daytime (Kumar et al., 2006).	**G** R	*High confidence: Likely* increases of HWDmax on annual time scale (T06; OS11). *Medium confidence:* Weaker signal than in most other land regions for HWDmax and weaker signal for HWDmean; less robust signal on seasonal time scale (OS11).	**G**	*Low confidence*: *slight or no increase in %DP10* (T06; OS11). *Low confidence:* More frequent & intense HPD over parts of S. Asia (Emori and Brown, 2005; Kamiguchi et al., 2006; Kharin et al., 2007; Rajendran and Kitoh, 2008).	**G** R	*Low confidence:* inconsistent signal of change in CDD and SM (T06; SW08; OS11).	**G**

	W. Asia	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11).	**G**	*High confidence: Likely* increases of HWD (T06; OS11).	**G**	*Low confidence*: Inconsistent signal of change in HP (T06; OS11).	**G**	*Low confidence:* inconsistent signal of change in CDD and SM (T06; SW08; OS11).	**G**
	Tibetan Plateau	*High confidence*: HD *likely* to increase & CD *likely* to decrease (OS11). *Medium confidence*: Large changes both in lower and higher quantiles of Tmax compared to other Asian regions but only for higher quantiles in winter (OS11).	**G**	*High confidence*: WN *likely* to increase (T06; OS11) and CN *likely* to decrease (OS11).	**G**	*High confidence: Likely* increases of HWD (T06; OS11).	**G**	*Medium confidence*: Increase of HP (T06; OS11).	**G**	*Low confidence:* inconsistent signal of change in CDD (T06; SW08; OS11).	**G**
Australia/New Zealand	Whole Australia and New Zealand	*High confidence*: HD *very likely* to increase & CD *very likely* to decrease in all regions (CSIRO and Bureau of Meteorology, 2007; Kharin et al., 2007; Mullan et al., 2008; OS11).	**G** **R**	*High confidence*: WN *very likely* to increase everywhere (T06; Kharin et al., 2007; Alexander and Arblaster, 2009; OS11) and CN *very likely* to decrease (OS11). *Medium confidence*: WN increase everywhere. Largest increases in WN in N compared with S and most consistent changes in inland regions (Alexander and Arblaster, 2009).	**G**	*High confidence: Likely* increase of HWDmax on the annual time scale (T06; OS11) and increase of HWD everywhere (Alexander and Arblaster, 2009). *Medium confidence:* Strongest increases in HWD in NW & most consistent increases inland (Alexander and Arblaster, 2009).	**G**	*Low confidence*: Lack of agreement regarding sign of change (T06; OS11). *Low confidence:* HPD tend to increase in E & decrease in W half of country – but considerable inter-model inconsistencies; HPC tends to increase everywhere – but considerable inter-model inconsistencies (Alexander and Arblaster, 2009).	**G**	*Lowto medium confidence:* Models agree on increase in CDD in southern Australia, but inconsistent signal over most of South Australia in SM; inconsistent signal in CDD and SM in Northern Australia (T06; SW08; OS11). Strongest CDD increases in western half of Australia (Alexander and Arblaster, 2009).	**G**
	Northern Australia	*High confidence*: HD *very likely* to increase & CD *very likely* to decrease (CSIRO and Bureau of Meteorology, 2007; OS11).	**G**	*High confidence*: WN *very likely* to increase (T06; Alexander and Arblaster, 2009; OS11) and CN *very likely* to decrease (OS11). *Medium confidence:* Changes larger than in Southern Australia (Alexander and Arblaster, 2009).	**G**	*High confidence: Likely* increases of HWDmax on the annual time scale (T06; OS11) and increases in HWD everywhere (Alexander and Arblaster, 2009). *Medium confidence:* Strongest increases in NW & most consistent increases inland (Alexander and Arblaster, 2009).	**G**	*Low confidence* (see whole region)	**G**	*Low confidence:* Inconsistent signal in CDD and SM (T06; SW08; OS11).	**G**
	Southern Australia and New Zealand	*High confidence*: HD *very likely* to increase & CD *very likely* to decrease (CSIRO and Bureau of Meteorology, 2007; OS11). *Low to medium confidence:* Strongest New Zealand increases in HD in North Island and largest decreases in frost days in South Island (Mullan et al., 2008).	**G** **R**	*High confidence*: WN *very likely* to increase (T06 , (T06; Alexander and Arblaster, 2009; OS11) and CN *very likely* to decrease (OS11). *Medium confidence:* Changes smaller than in Northern Australia (Alexander and Arblaster, 2009).	**G**	*High confidence: Likely* increases of HWDmax on the annual time scale (T06; OS11) and increase in HWD everywhere (Alexander and Arblaster, 2009). *Medium confidence:* Most consistent increases inland (Alexander and Arblaster, 2009).	**G**	*Low confidence* (see whole region) *Low to medium confidence:* In New Zealand, increase in HP events at most locations (Mullan et al., 2008; Carey-Smith et al., 2010).	**G** **R**	*Medium confidence:* Models agree on increase in CDD in southern Australia including SW (T06; Alexander and Arblaster, 2009; OS11), but inconsistent signal in SM over most of the region, slight decrease in SW (T06; SW08; OS11).	**G**

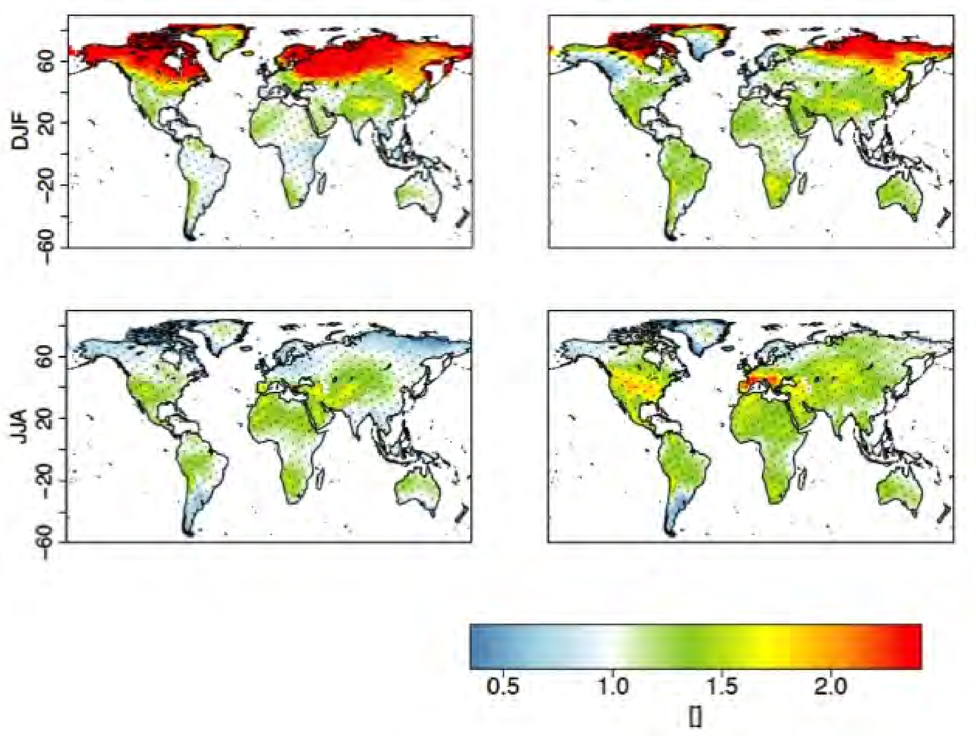


Figure 3.1: Scaling between globally-averaged annual mean projected change in Tmax and spatial changes in seasonal (DJF, top; JJA, bottom) changes in 10%ile (left) or 90%ile (right) of Tmax, CMIP3 projections, 2080-2100 time frame minus 1980-2000 time frame (A2 vs 20C3M). The 10%ile and 90%ile values are computed from pooling all data for the respective months in the two 20-year periods. [adapted from Orlowsky and Seneviratne, 2011]

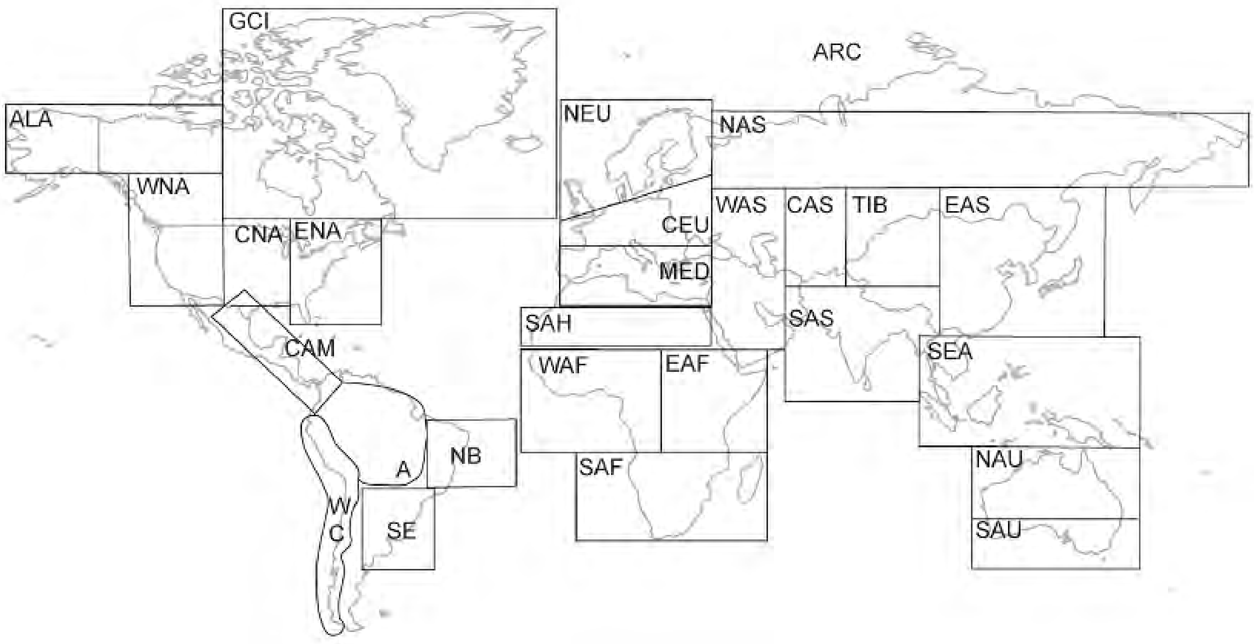


Figure 3.2: Definitions of regions used in Tables 3.2 and 3.3.

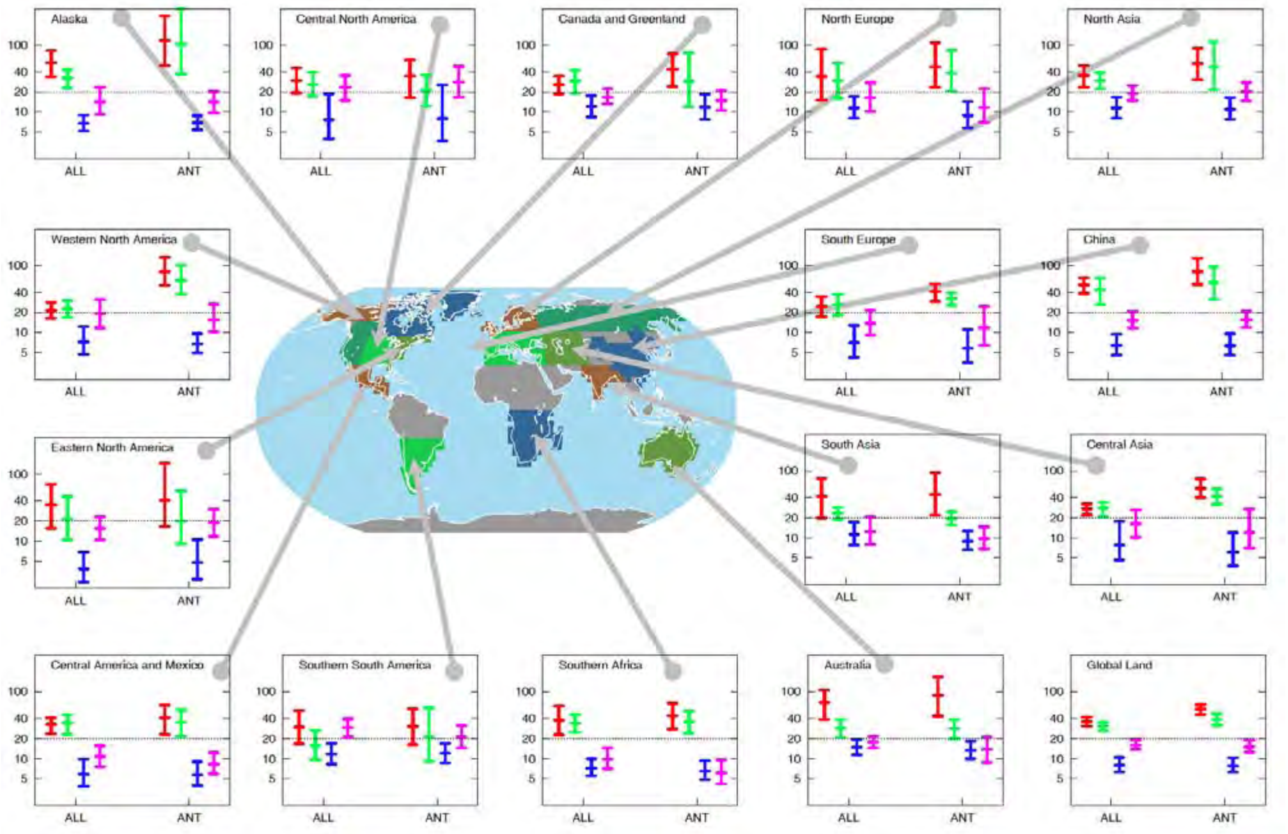


Figure 3.3: Estimated waiting time (years) and their 5% and 95% uncertainty limits for 1960s 20-yr return values of annual extreme daily temperatures in the 1990s climate (see text for more details). From Zwiers et al., (2011). Red, green, blue, pink error bars are for annual minimum daily minimum temperature (TNn), annual maximum daily minimum temperature (TNx), annual minimum daily maximum temperature (TXn), and annual maximum daily maximum temperature (TXx), respectively. Grey areas indicate insufficient data.

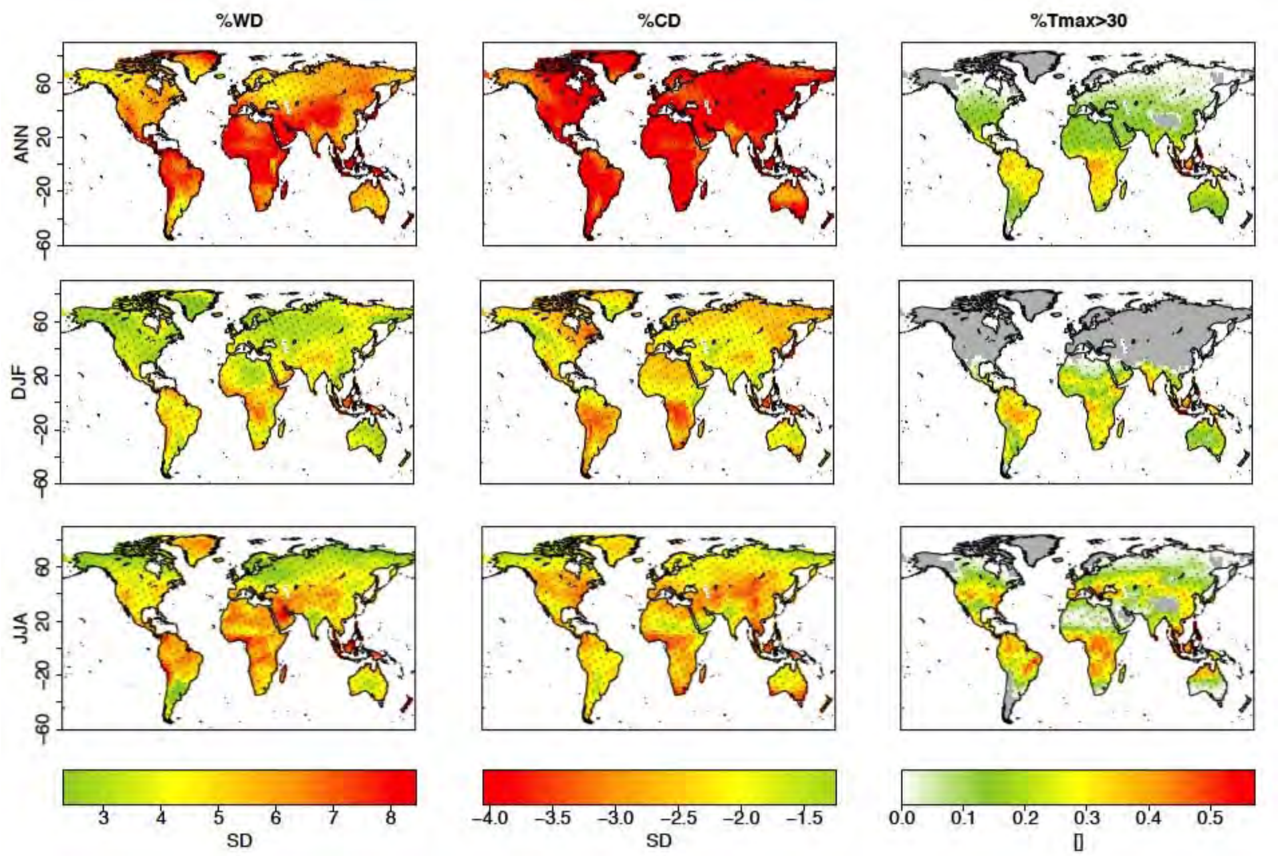


Figure 3.4: Projected annual and seasonal changes of three indices for Tmax: Fraction of warm days, fraction of cold days, and fraction of days with Tmax > 30°C; CMIP3 projections, 2080-2100 time frame minus 1980-2000 time frame (projections for A2 scenario, relative to late 20th century (20C3M) simulations), annual (top), DJF (middle) and JJA (bottom). Shading is only applied for areas where at least 66% of the models agree in the sign of the change; stippling is applied for regions where at least 90% of all models agree in the sign of the change [from Orlowsky and Seneviratne, 2011, after Tebaldi et al., 2006].

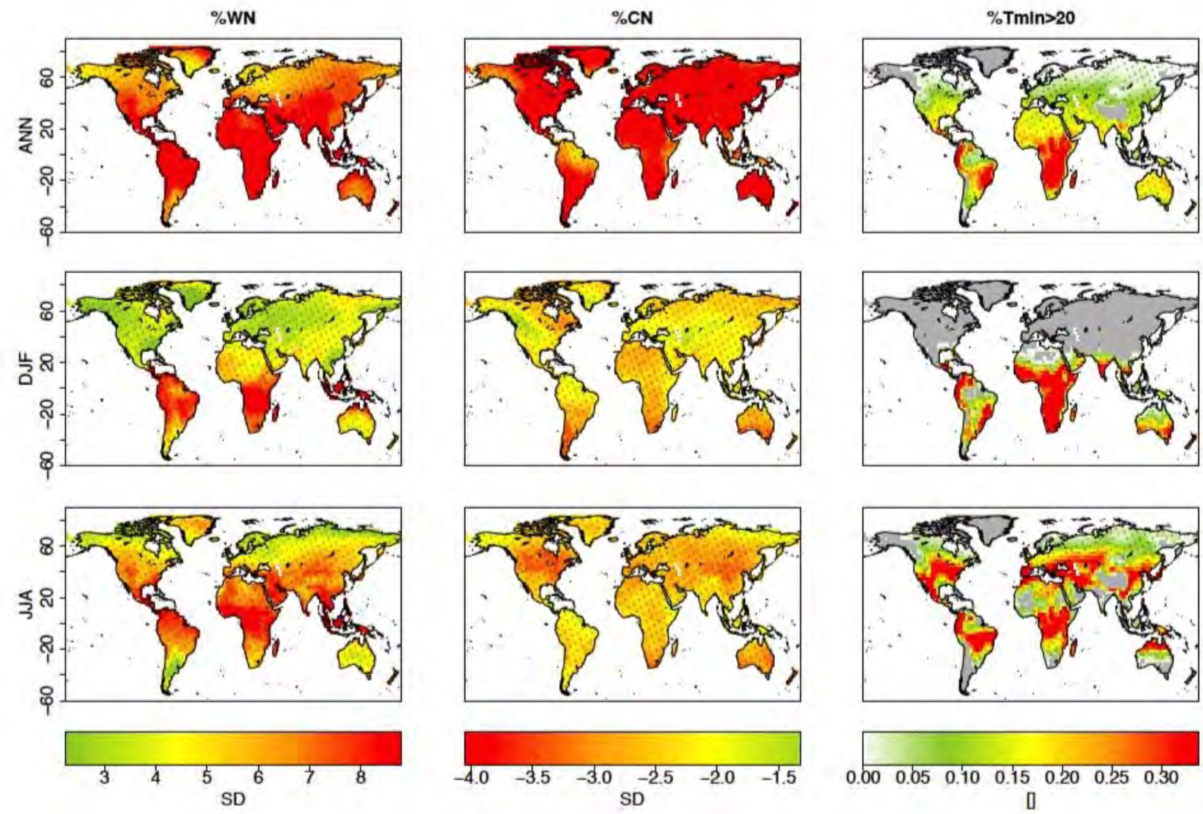


Figure 3.5: Projected annual and seasonal changes of three indices for Tmin: Fraction of warm nights, fraction of cold nights, and fraction of days with Tmin > 20°C; CMIP3 simulations, 2080-2100 time frame minus 1980-2000 time frame (projections for A2 scenario, relative to late 20th century (20C3M) simulations), annual (top), DJF (middle) and JJA (bottom). Shading is only applied for areas where at least 66% of the models agree in the sign of the change; stippling is applied for regions where at least 90% of all models agree in the sign of the change [from Orlowsky and Seneviratne, 2011, after Tebaldi et al., 2006].

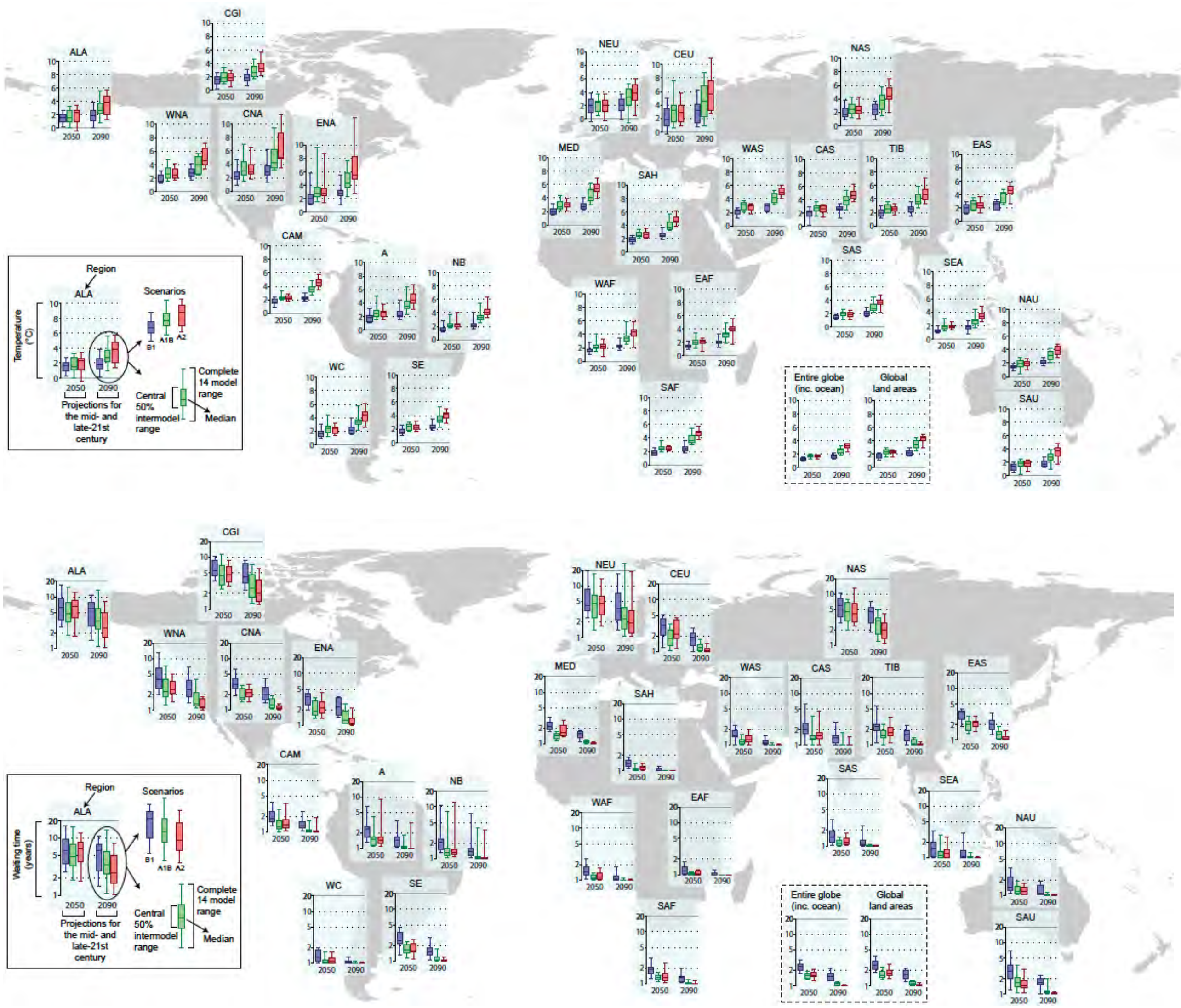


Figure 3.6: (a, top) Projected changes from the late-twentieth-century 20-year return values of annual maximum of the daily maximum temperature in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the CMIP3, under three different SRES emission scenarios B1 (blue), A1B (green) and A2 (red); units in °C. Adapted from the analysis in Kharin et al. (2007). The vertical extent of the whiskers shows the range of projected changes from all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Model projections suggest that the the 20-year extreme annual daily maximum temperature will increase by about 2°C by mid-21st century and by about 4°C by late-21st century, depending on the region.
(b, bottom) Projected waiting times for late-twentieth-century 20-year return values of annual maximum of the daily maximum temperature in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the CMIP3, under three different SRES emission scenarios B1, A1B and A2 Adapted from the analysis in Kharin et al. (2007). The vertical extent of the whiskers shows the range of projected changes from all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Model projections suggest that the waiting time for a late 20th century 20-year extreme annual daily maximum temperature will be reduced to about 2-20 years by mid-21st century and by about 1-5 years by late-21st century, depending on the region. Two global domains for which projections are shown are: the entire globe including the oceans, and the global land areas.

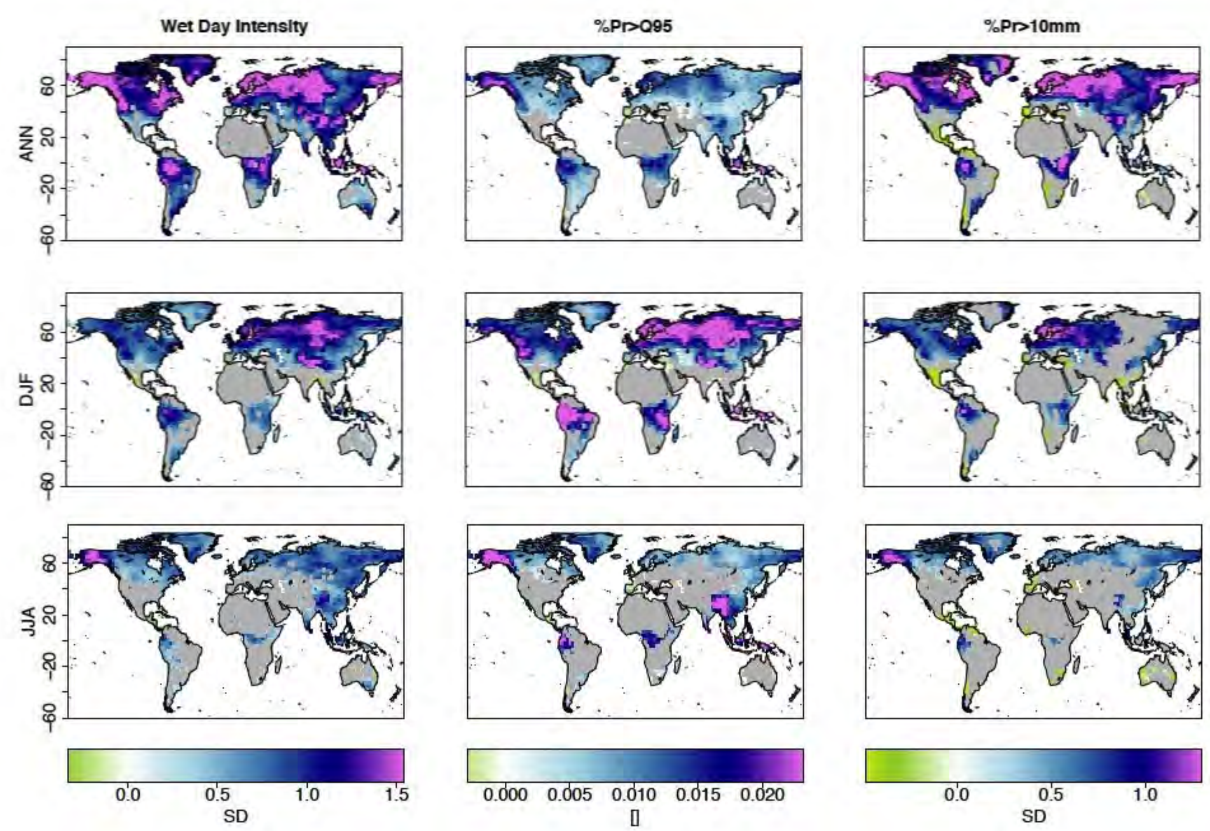


Figure 3.7: Projected annual and seasonal changes of three precipitation indices: Wet day intensity, fraction of days with precipitation above the 95%-quantile of daily wet day precipitation and fraction of days with pr > 10 mm; CMIP3 simulations, 2080-2100 time frame minus 1980-2000 time frame (projections for A2 scenario, relative to late 20th century (20C3M) simulations), annual (top), DJF (middle) and JJA (bottom). Shading is only applied for areas where at least 66% of the models agree in the sign of the change; stippling is applied for regions where at least 90% of all models agree in the sign of the change [from Orlowsky and Seneviratne, 2011, after Tebaldi et al., 2006].

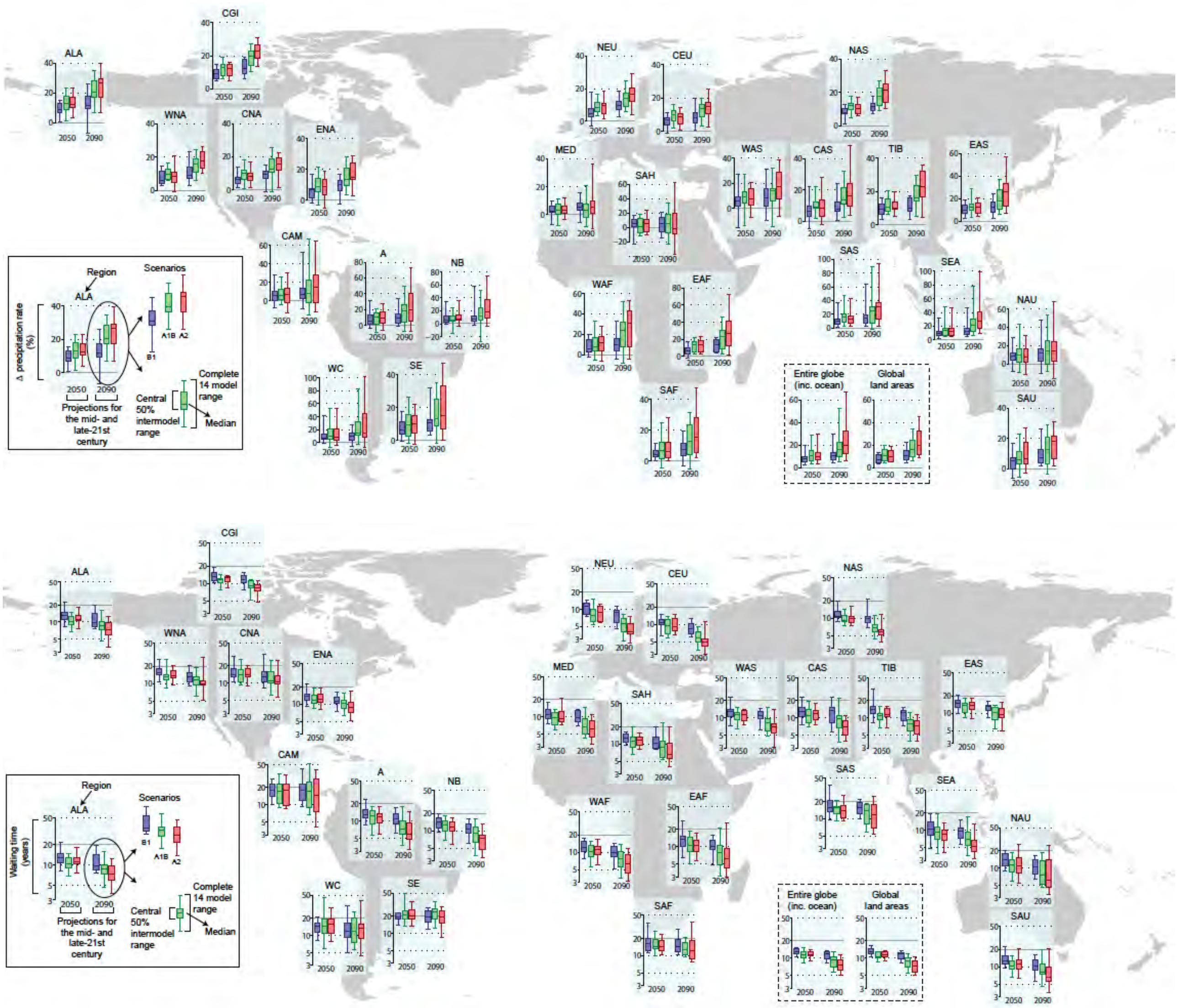


Figure 3.8: (a, top) Projected changes from the late-twentieth-century 20-year return values of annual maximum 24-hour precipitation rates (%) in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the CMIP3, under three different SRES emission scenarios B1 (blue), A1B (green) and A2 (red); (adapted from Kharin et al., 2007). The vertical extent of the whiskers shows the range of projected changes from all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Although the uncertainty range of projected change in extreme precipitation is large, the median model projection is that the extreme 24-hour precipitation rate will increase by about 5-10% by mid-21st century and by about 10-20% by late-21st century, depending on the region and the emsissions scenario.

(b, bottom) Projected waiting times for late-twentieth-century 20-year return values of annual maximum 24-hour precipitation rates in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the CMIP3, under three different emission scenarios SRES B1 (blue), A1B (green) and A2 (red) (adapted from Kharin et al., 2007). The vertical extent of the whiskers in both directions describes the range of projected changes by all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Although the uncertainty range of projected change in extreme precipitation is large, almost all models suggest that the waiting time for a late 20th century 20-year extreme 24-hour precipitation event will be reduced to substantially less than 20 years by mid-21st and much more by late-21st century, indicating an increase in frequency of the extreme precipitation at continental and sub-continental scales under all three forcing scenarios. Two global domains for which projections are shown are: the entire globe including the oceans, and the global land areas.

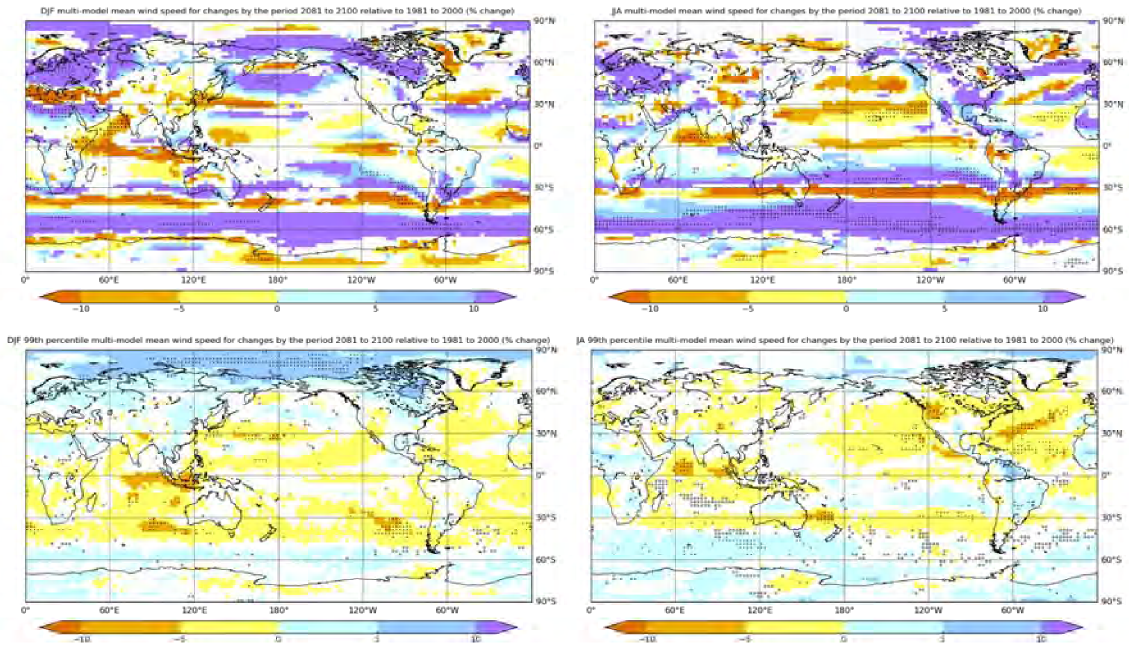


Figure 3.9: The average of the projected multi-model 10 m mean wind speeds (top) and 99th percentile daily wind speeds (bottom) for the period 2080 to 2099 relative to 1980 to 1999 (% change) for December to February (left) and June to August (right) plotted only where more than 66% of the models agree on the sign of the change. Fine black stippling indicates where more than 90% of the models agree on the sign of the change and bold grey stippling (in white or light coloured areas) indicates where 66% of models agree on a small change between ±2%. From McInnes et al., (2011).

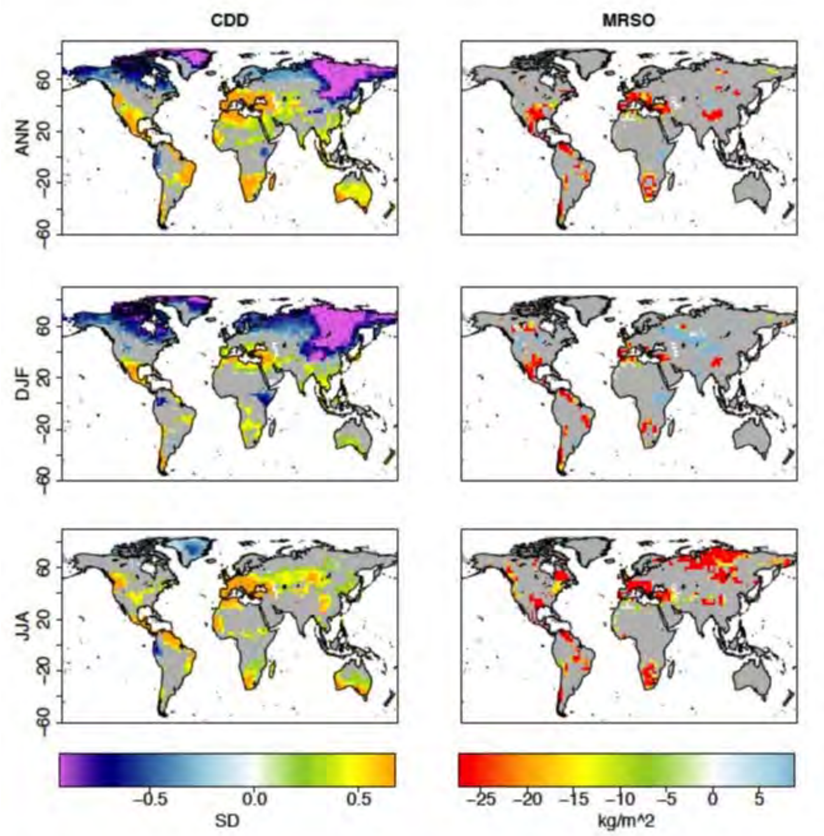
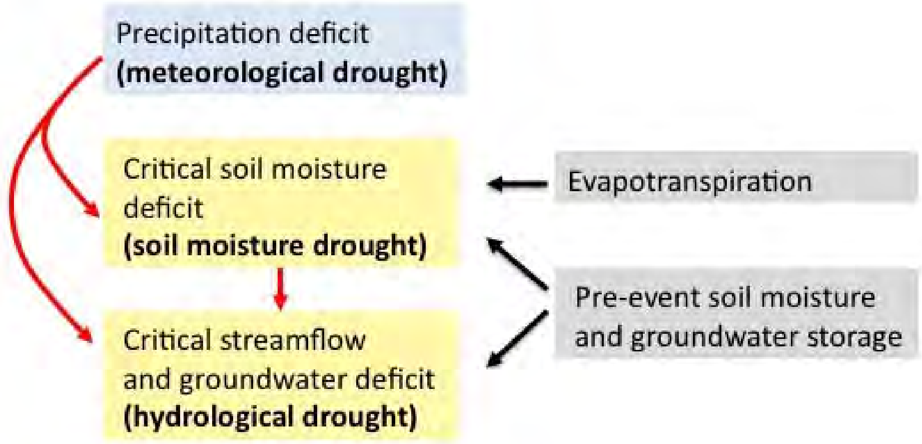
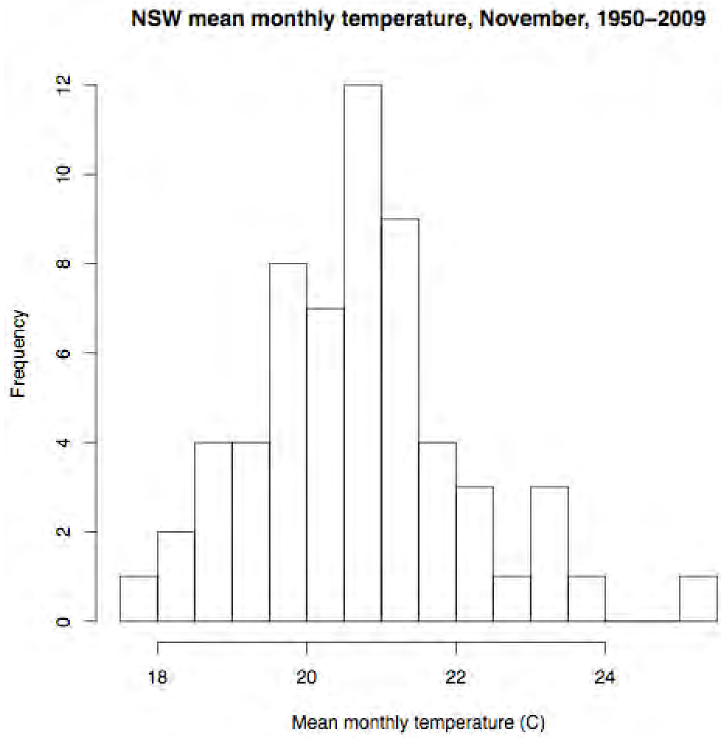


Figure 3.10: Projected annual and seasonal changes of two dryness indices: Consecutive dry days (CDD, days with pr < 1 mm) and average soil moisture (mrso); CMIP3 projections, 2080-2100 time frame minus 1980-2000 time frame (A2 relative to 20C3M simulations), annual (top), DJF (middle) and JJA (bottom). Shading is only applied for areas where at least 66% of the models agree in the sign of the change; stippling is applied for regions where at least 90% of all models agree in the sign of the change [from Orlowsky and Seneviratne, 2011, after Tebaldi et al., 2006].



Box 3.2, Figure 1: Processes and drivers relevant for meteorological, soil moisture, and hydrological droughts.



FAQ 3.2, Figure 1: The distribution of monthly mean November temperatures averaged across the State of New South Wales in Australia, using data from 1950–2009. Data from Australian Bureau of Meteorology. The mean temperature for November 2009 (the bar on the far right hand end of the figure) was more than three standard deviations from the long-term mean (calculated from 1950–2008 data).